Doctoral Thesis

Climate Change Detection in Central Africa

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Date of Defence

14.11.2007

Statutory Declaration

I herewith declare that this thesis is the result of my original work carried out at Brandenburg University of Technology, Cottbus, Germany within the framework of the Doctoral program in Environmental and Resource Management. My research was supervised by Prof. Dr. rer.nat. Schaller and Prof. Dr-Ing. Dr. h.c. Michael Schmidt, BTU, Cottbus, Germany.

All the work within this text is original except where cited from an additional source. Quotations and other indirect information have been clearly marked and noted. This work has not been submitted previously or to other examination bodies, and is currently unpublished.

Nana Takyi Wilberforce
Cottbus, 17th November 2007

Abstract

Central Africa is a band of land stretching from the Atlantic Ocean to the Horn of Africa between 18 º N 18 º W and 0.75 º S 44 º E. The climate of this region shows a transition zone from the Sahara Desert in the north to the wet tropical Guinean coast and equatorial continental interior in its south. In such transition zones, drought is a common occurrence; of great concern is the progressive drying of the region from the 1970s and into the 1980s. As documented by Tucker et al. (1991), the area of the Sahara grew from about 8.6 to 10 million km from 1980 to 1984. But in the following year 1985 which was a year of stronger rainfall, the area decreased again by 0.7 million km. More recently, rainfall levels over Central Africa have increased to the point of causing regional flooding in 2005. But the higher rainfall regime of the pre-1960s has not returned as yet in Central Africa. This thesis aims to investigate whether climate change has occurred in Central Africa in the 2nd half of the 20th century. The reanalysis data from the National Centres for Environmental Prediction (NCEP) was used to analyse trends of temperature and precipitation over Central Africa with the linear regression model and patterns of precipitation using principal component analysis. The results of the trend analyses revealed an average increase in temperature of 0.15K/ decade and an average decrease in precipitation of -91 mm/decade for the period 1948-2004 over the whole study area. A regional trend analysis of box averaged data (2.5° in latitude and longitude) revealed an increased precipitation over the Northern part," the Sahel", and a marked decrease in precipitation and an increase in temperature over the Eastern parts of the study areas especially in countries such as Somalia, Sudan and Ethiopia. To quantify the changes in temperature and precipitation for different climatic periods (1951-1980, 1961-1990 and 1971-2000), 30-year trend analyses were also performed. Results revealed a relatively stable temperature trend and fluctuations in the precipitation trend over the three 30 year climate periods.

Principal component analysis produces a decomposition of the data field into spatial patterns (eigenvectors) and a temporal time series describing the temporal importance of these patterns. For precipitation over Central Africa, three patterns stand out for 60.39% of the total variance in the precipitation over Central Africa. The individual principal components explain 37.6%, 12.1%, and 10.7% respectively of the total variance. The first principal component captures progressive drying in Central Africa, the second principal component captures the different seasons within Central Africa and the third principal component captures the movement of the Intertropical Convergence Zone over Central Africa. Apart from explaining the current precipitation patterns over Central Africa, the use of principal component analysis in this study has demonstrated that a limited number of spatial patterns are basis for African weather, coordinate system for present day climate with maximum variance contribution along the first axis, maximum of the remaining variance along the second axis with subsequent axes explaining less variance can be used to explain precipitation variability within Central Africa. The same method can be used to explain temporal differences between climate change scenarios and present day climate.

Zusammenfassung

Zentralafrika erstreckt sich vom Atlantik bis zum Horn von Afrika zwischen 18 º N 18 º W und 0.75 º S 44 º E. Das Klima dieser Region geht von der trockenen Sahara im Norden in die feuchte tropische Küste über. In solchen Übergangszonen, besonders im kontinentalen Inneren, tritt häufig Dürre auf. Von großer Bedeutung war dabei eine zunehmende Austrocknung der Region während der siebziger und achtziger Jahre. Wie von Tucker et al. (1991) dokumentiert, dehnte sich die Sahara 1980-1984 von ungefähr 8.6 auf 10 Millionen Quadratkilometer aus. Aber im folgenden Jahr 1985, in dem wieder mehr Niederschlags fiel, schrumpfte die Wüste um 0.7 Millionen Quadratkilometer. Vor kurzem (2005) führten Niederschlagsüberschüsse sogar zu regionalen Überschwemmungen. Aber das höhere Niederschlagsniveau der Zeit vor 1960 hat bis jetzt in Zentralafrika noch nicht wieder eingestellt.

Diese Arbeit zielt darauf ab zu erforschen, ob eine Klimaveränderung in Zentralafrika in der 2. Hälfte des 20. Jahrhunderts eingetreten ist. Die Reanalyse-Daten des National Centre for Environmental Prediction (NCEP) wurden verwendet, um Tendenzen der Temperatur und des Niederschlags mit einer linearen Regression zu analysieren. Niederschlagsmuster wurden mit der Hauptkomponentenanalyse (principal component analysis) untersucht. Die Resultate der Trendanalysen ergaben eine durchschnittliche Zunahme der Temperatur um 0.15 K/Dekade und eine durchschnittliche Abnahme des Niederschlags um -91 mm/Dekade für den Zeitraum 1948-2004 über das gesamte Untersuchungsgebiet. Eine regionale Trendanalyse zeigte erhöhten Niederschlag über dem Norden und der Sahelzone, und eine deutliche Abnahme über den östlichen Gebieten, die mit einem Anstieg der Temperatur in Ländern wie Somalia, Sudan und Äthiopien einherging. Um die Veränderungen von Temperatur und Niederschlag für unterschiedliche klimatische Perioden (1951-1980, 1961-1990 und 1971-2000) zu beschreiben, wurden 30-Jahr Trendanalysen durchgeführt. Die Resultate zeigten eine verhältnismäßig beständige

Temperaturtendenz, aber Fluktuationen beim Niederschlag. Die Hauptkomponentenanalyse ermöglicht eine Aufspaltung der Daten in räumliche Muster (Eigenvektoren) und eine Zeittreihe, die zeitliche wichtigkeit dieser Muster beschreibt. Für den Niederschlag in Zentralafrika stechen drei Muster heraus, die 60,4 % der Gesamtabweichung repräsentieren. Die einzelnen Muster erklären dabei 37,6 %, 12,1 % und 10,7 % der Gesamtabweichung. Das erste Muster beschreibt Austrocknung, das zweite die unterschiedlichen Jahreszeiten, und das dritte die Bewegung der Intertropischen Konvergenz Zone (ITCZ). Die Hauptkomponentenanalyse erklärt nicht nur die gegenwärtigen Niederschlagsmuster in Zentralafrika, sondern zeigt, dass eine begrenzte Anzahl von räumlichen Mustern Grundlage für das afrikanische Wetter sind. Sie beschreibt ein Koordinatensystem für das heutige Klima mit maximalen Abweichungen entlang ersten Achse und maximalen restlichen Abweichungen entlang der zweiten Achse, während die folgenden Achsen geringere Abweichungen erklären und verwendet werden können um die Niederschlagveränderlichkeit in Zentralafrika zu bestimmen. Die gleiche Methode kann verwendet werden, um zeitliche Unterschiede zwischen Szenarien der Klimaänderung und dem heutigem Klima zu erklären.

DEDICATION

This work is dedicated to God Almighty who is the Author and Finisher of our faith, to my understanding caring and loving wife maame abena and lastly to my parents and my two wonderful brothers ato and his disciple (apostle john) grandpa.

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Dr Christen and Dr Sungman provided assistance with elements of the computer programming, particularly in the early stages of the project, and are thanked for their patience and the provision of certain IDL scripts. Naana Adoma Boampong, Doris Zama and Techoro Prosper proofread and commented on various chapter drafts.

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1 Introduction and Outline

Climate change is often synonymous with global warming to the general public, the media or policy makers. This is no surprise because the reported increase in mean temperature near the surface of the earth provides the most apparent information to society that our climate is changing. The term "climate change" is sometimes used to refer to all forms of climatic inconsistency, but because the Earth's climate is never static, the term is more properly used to imply a significant change from one climatic condition to another. Scientists however, tend to use the term in the wider sense to also include natural changes in climate, they refer to climate change as a statistically significant variation in either the mean state of the climate or in its variability, persisting for an extended period of time, normally decades or longer (Houghton et al. 2001). Thus climate change is not only a change in temperature but also a change in other variables in the climate system such as precipitation.

The effects that a global change in mean temperature or precipitation may have on local and regional scales are extremely variable and uncertain due to the influence of atmospheric circulation and ocean bodies (Mitchell and Hulme, 1999). In addition to changes in the mean climatic conditions (such as temperature and rainfall), the frequencies of irregular seasons and extreme events, including fires, hurricanes and droughts, are likely to change and in some places increase (Parry, 1986; Peters, 1992). Given the potentially dramatic effects on local climate, natural resources, infrastructure and economic activities, Africa may be particularly vulnerable to and at risk from climate change. Africa shows one of the most sensitive ecological transitions on the globe i.e. from the Northern deserts to the tropical rain forests. This transition is likely to change because of the global climate change. The transition region in Africa is described in this study as Central Africa and includes the Sahel, Sudan, N & S Guinea and Rainforest. Figure 1 presents annual rainfall anomalies for the entire twentieth century. The anomalies are aggregated for the entire Sahel region, 12-18°N latitude, which is part of the study area. The anomalies are calculated as follows: The mean annual observed rainfall representing the spatially aggregated Sahel over the period 1901-1996 is calculated to be 467 mm. This compares with a maximum value of 608 mm in 1950, a minimum value of 340 mm in 1984. The standard deviation (s) of the annual rainfall over the entire period is 59 mm. The minimum and maximum anomalies correspond to deviations from the mean of -2.22 s and 2.41 s , or -32% and +31%, respectively. Years after 1968 are characterised by negative anomalies, and the last three decades of the twentieth century represent a persistently more arid regime than that prevailing throughout the earlier times of the century. Since the 1970s there has been some concern about aridification of Central Africa's environment. In the Sahel for example successive rainfall shortages have resulted in widespread droughts in 1972 and 1973. Dyson (1975) suggested the region may be experiencing a move towards an arid climate regime and even though the region recovered from the early droughts, rainfall amounts continue to dwindle. The largest rainfall shortage was experienced in 1984 with rainfall again low in 1985 and 1986 disproving the suggestion the drought will end in 1985 (Fraure et. al. 1981).

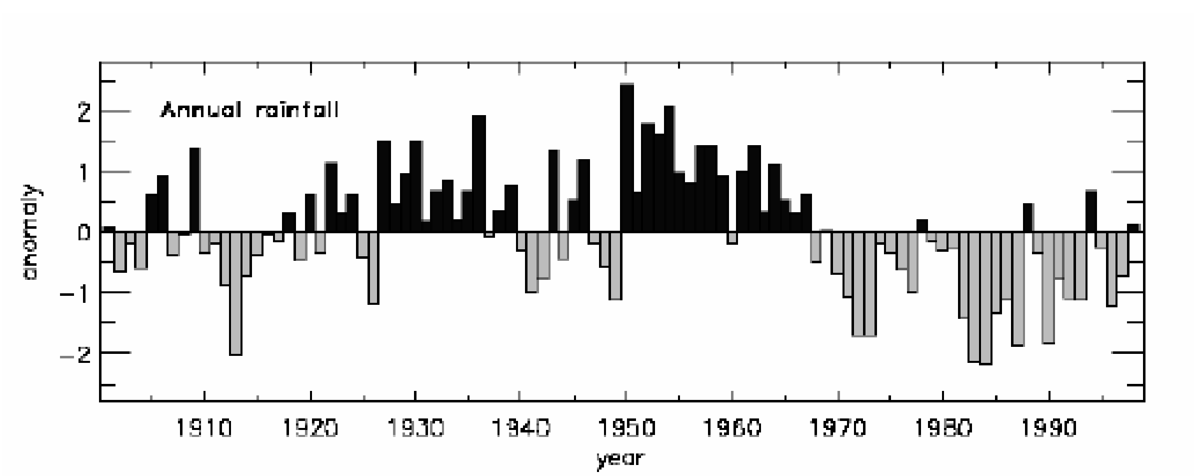


Figure 1: Annual rainfall anomalies for the Sahel, anomalies are calculated from the mean for the entire series (1901-1998) from the dataset of New et al. (2000).

1988 was a wetter year; however, rainfall anomalies remained negative till 1993. The period from 1995-1997 was dry with 1998 showing values close to the century mean. Many authors have described the period since the late 1960s as representing a time noted for its constant drought. Hulme (1996) describes this recent desiccation as contributing to an equivalent mean linear trend amounting to a 21% decrease in rainfall over the twentieth century. Climate change, drought and desertification remain linked because of the interplay in temperature and precipitation. Therefore, a detailed analysis of climate change occurring in a region is an important precursor for any measure to combat drought and desertification within that region.

From a scientific point of view, assessing changes in climate in Africa is problematic because the datasets that are available to study past climates are biased towards recent years simply because of improved handling of technology, the installation of automatic stations and improved technical know-how of observation technicians. Substantial work has been completed in recent years in the area of climate trend research in Central Africa, but in the absence of complete and long term datasets, the regional trend studies needed to rely on combining the results of different analysis from different areas within Central Africa. Unfortunately, the exact definition of the measures that were used to characterize the past climatic trends in terms of the period of analysis taking into consideration the different seasonal periods over various local regions in Central Africa differed from one study to the other. To gain a uniform perspective on changes in past climates, a homogeneous and long-term climate dataset should be used for analysis. This study aims to update information on changes in climate over Central Africa through the analysis of the reanalysis data from the National Centres for Environmental Prediction (NCEP) and to identify patterns of Central African precipitation.

The study addresses key questions such as:

1. What are the observed trends in temperature and precipitation over Central Africa in the 2nd half of the 20th century?

2. Which data sources are available that can be used in order to investigate the 1st question?

3. Is there an identifiable pattern in rainfall variations over Central Africa?

The contents of all the chapters are summarized as follows: Chapter 2 is a combination of review, synthesis and discussion of research carried out by many authors. The work discussed largely relates to research into mechanisms of Central African rainfall modulation. Chapter 3 addresses the issue of data scarcity in Central Africa and describes some data sets available for climate research in Central Africa. The methods of analyses as applied in this study are also discussed in this chapter.

Chapter 4 presents results and discussion from the analyses of the temperature and precipitation datasets using the methods described in chapter 3. Chapter 5 is a synthesis of the research presented in previous chapters, the findings of this thesis, conclusions and recommendations for further research.

2 The scientific context of climate change in Central Africa

2.1 Introduction

The term 'Central Africa' is used in this study to imply a region between upper left corner 18 º N 18 º W and lower right corner 0.75 º S 44 º E, see Figure 2. It ranges from the West African coast through the Sudan, highlands of Ethiopia to the horn of Africa shifting from semi arid grassland, humid equatorial to thorny savannah and receives roughly 150-2500 mm of rainfall each year mainly in the monsoon season.

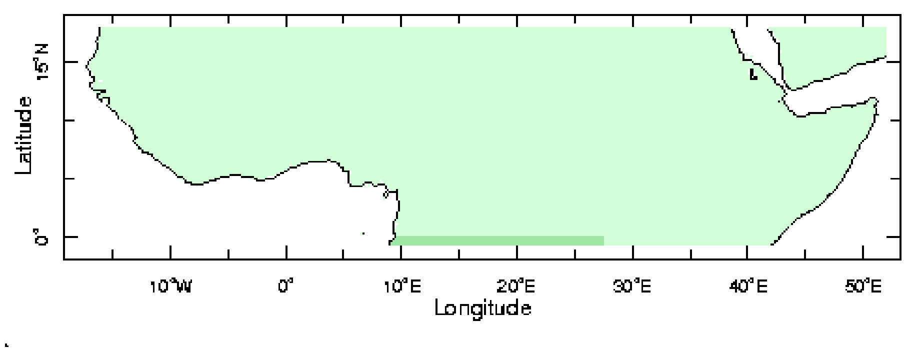


Figure 2: Map of Central Africa

Much of the work reviewed has been carried out as a direct result of a desire to understand processes related to climate change in Central Africa. The development of ideas concerning mechanisms of rainfall modulation is discussed in Section 2.2. Section 2.3 addresses mechanisms of rainfall modulation in Central Africa. The research into changes in global temperature patterns and their correlations with Central African rainfall is summarized, with particular attention being paid to sea surface temperature patterns and variations in the inter-hemispheric temperature contrast. The impact of sulphate and other aerosols is discussed briefly in this section, due to the possible effect of such particles on hemispheric-scale temperature patterns, which are associated with variations in Central African rainfall. Relationships between Central African rainfall and global scale atmospheric circulation are broadly reviewed.

2.2 Development of ideas concerning mechanisms of rainfall modulation

The nature of the recent period of reduced rainfall in Central Africa has been discussed in Chapter 1. The duration of this dry episode has given rise to concern that the Central African region may be experiencing a “climatic step” towards a more arid regime (Kidson, 1977; Rognon, 1987). It was speculated as early as 1975 that aridification in Central Africa might represent a new mean regional climate, perhaps as the result of anthropogenic influences on the global atmosphere or regional land-atmosphere interaction (Dyson-Hudson and Dyson-Hudson, 1975; Kelly, 1975; Newman, 1975). Such concerns have encouraged various authors (Nicholson, 1976, 1978) to investigate the historical precedents for such a dry episode. However, such studies suffer from the scarcity of historical and palaeoclimatic data, as well as from the poor temporal resolution inherent in these data and the difficulty in quantifying changes in rainfall before the beginning of the instrumental period.

Early studies of rainfall changes in arid and semi-arid regions such as Central Africa emphasised the potential of changes in vegetation cover to lead to changes in the processes governing heat and moisture exchanges between the land and atmosphere (Charney et al., 1975). Such studies assumed that changes in ground cover, arising from drought and/or the removal of vegetation by humans and animals, would be widespread enough to cause changes in the atmospheric heat and moisture budgets on a large scale. However, many authors have regarded the presumed extent of such hypothetical changes as unrealistic (Hulme and Kelly, 1993; Mortimore, 1998). Indeed, assumptions of widespread land degradation throughout Central Africa are largely unfounded, as data are available for only a limited number of locations that are not necessarily representative of the region as a whole (Mace, 1991; Stocking, 1996; Williams and Balling, 1996).

Other early investigations into the recent period of desiccation associated dry conditions in Central Africa with large-scale atmospheric circulation patterns (Kidson, 1977; Lamb, 1978). Kidson (1977) linked drought-related changes in the regional circulation with global atmospheric changes. Such large-scale changes in the atmospheric circulation appear to have persisted throughout much of the period of desiccation (Shinoda, 1990). The implication of this is that the multi-decadal dry episode witnessed recently in Central Africa is at least partly the result of a (perhaps temporary) shift in the global circulation. Associated with such atmospheric circulation changes are changes in patterns of sea surface temperatures (SSTs). Many investigations have concentrated on the association of dry years in Central Africa with particular modes of SST variation both within particular ocean regions and on a hemispheric and global scale (Folland et al., 1986; Ward, 1998). The scenario in which drought in Central Africa is linked with changes in atmospheric circulation and particular SST patterns is supported by work such as Newell and Hsiung (1987), Street-Perrott and Perrott (1990), Fontaine and Bigot (1993), who report different modes of SST variation and changes in wind fields before and after the onset of the dry episode. It may also be hypothesised that the onset of dry conditions was due to large-scale atmosphere-ocean dynamics, but that the continuation of drought conditions has been due to regional feedback processes, which will continue to reinforce drought conditions until a large-scale circulation anomaly occurs which is of sufficient magnitude and of the appropriate nature (i.e. opposite in sense to that which triggered the dry conditions) to overcome the feedback processes and re-establish a wetter regime (Nicholson, 1995; Shukla, 1995). As discussed in Chapter 1, annual rainfall in Central Africa exhibits far more persistence after 1970 than in the previous half of the twentieth century (Nicholson, 1995; for a further discussion, see Chapter 4). This suggests that either the mechanisms modulating Central African rainfall are less variable on a year-to-year timescale in the 2nd half of this century compared to the former and regional feedback processes have acted so as to reinforce drought conditions, or a different regional or global climatic regime acts in concert with regional feedback mechanisms to sustain the dry episode.

2.3 Mechanisms of Central African rainfall modulation

2.3.1 Central African climate change and the land surface

At the beginning of the period of twentieth century desiccation, many authors attempted to explain the observed changes in the Central African climate in terms of land-atmosphere interaction. The prevailing notion was that widespread changes in the land surface had resulted in changes in the forcing of the regional atmosphere that led to conditions favourable for enhanced aridity. The most well-known of these theories was the biogeophysical feedback model (Charney et al., 1975). In this model, removal of vegetation by human activity such as deforestation and livestock grazing increases the albedo of the land surface and leads to a reduction in the radiative heating of the atmosphere, causing enhanced subsidence and reducing the potential for convective rainfall. This model was supported by modelling studies (Charney et al., 1977; Sud and Fennessy, 1982; Laval and Picon, 1986). The removal of vegetative cover would result in increased wind and water erosion of soils, leading to land degradation and desertification (Rapp, 1986; for a discussion, definitions of land degradation, see Imeson (1991) and Mortimore (1998)).

Such degradation would reduce the capacity of the soil to hold moisture, and the postulated subsequent reduction in available surface moisture would also reduce the likelihood of rainfall, particularly from the squall lines which bring most of the rainfall to Central Africa, and whose development is partly dependent on surface moisture sources (Rowell and Milford, 1993). The importance of these processes was also supported by modelling studies (Sud and Fennessy, 1984; Cunnington and Rowntree, 1986).

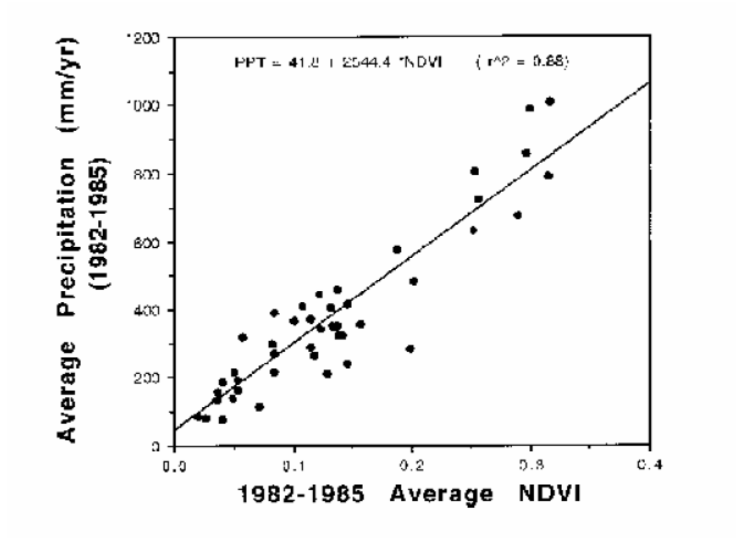
The above explanations for the Central African drought rested on the assumption of widespread and significant land degradation and desertification. The idea that the Central African environment had deteriorated in the manner described above was well established in the 1970s and 1980s, and an influential paper by Lamprey (1975) claimed that the Sahara had advanced by some 90km to 100 km in the north Kordofan region of Sudan between 1958 and 1975. Lamprey also cited "…ecological degradation….largely due to past and current land use practices….accelerated during periods of drought." Sand encroachment was "….the result of several thousand years of abuse of the fragile ecosystems which formerly existed in the Sahara and Nubian areas." He concluded that there was a "….need to educate the rural population, particularly as many of the problems are due to traditional and hitherto unquestioned practices."

Lamprey's conclusions echoed those of colonial authors. The notion of an advancing Sahara originated in the 1930s and 1940s, when many European and North American observers interpreted seasonal and interannual changes in Central African vegetation as being indicative of a southward expansion of the Sahara (For a detailed discussion and critique of theories of desertification and land degradation see Mortimore (1998)). Such changes are generally a result of natural variability in the regional climate of Central Africa - high variability in rainfall and hence in the physical environment are a feature of semi-arid areas in general (Thomas, 1993, 1997). On an interannual timescale in Central Africa, vegetation bands will shift northwards and southwards during periods of high and low rainfall respectively. The dynamic nature of the geomorphological processes operating in the region means that blown sand and mobile dunes will penetrate southward under dry conditions. However, such shifts in climatic and ecological zones are as likely to represent an oscillation of the so-called "desert boundary" (Tucker et al. 1991) in response to short-term changes in rainfall as they are to represent a steady and systematic southward encroachment of the Sahara. Studies of vegetation cover using the Normalised Difference Vegetation Index (NDVI) have shown that vegetation quickly recolonises “desertified” regions when rainfall permits (Hess, 1996; Tucker et al. 1991). Whereas Lamprey (1975) interpreted the different locations of the “desert boundary” in 1958 and 1975 to be the result of a systematic expansion of the Sahara, Tucker et al. (1991, 1994) have demonstrated that the location of this boundary (defined in terms of vegetation cover) varies from year to year in association with rainfall amounts. Figure 3 shows wet-season NDVI, averaged over Central Africa for the years 1980 to 1994. The NDVI is a measure of vegetation cover, and is well correlated with rainfall. The decline in NDVI until 1984, and the subsequent rapid recovery of NDVI values, demonstrates that vegetation quickly recolonised areas affected by dry conditions in the early 1980s, suggesting that the impact of successive dry years on the land surface is reversible. A study by Helldén (1988) also failed to find any evidence of severe desertification in north Kordofan over the period investigated by Lamprey (1975), but did demonstrate a short-term impact of drought on the vegetation in the region. Nicholson and Tucker (1998) state that there has been “…no progressive change of either the Saharan boundary or vegetation cover in Central Africa…, nor has there been a systematic reduction of ‘productivity’ ” between the early 1980s and late 1990s. Over the same period they suggest little change in surface albedo has occurred, although they claim that a change in albedo of up to 10 % since the 1950s is “conceivable”. This should be compared with a maximum seasonal variation between July 1989 and September 1990 of 6.5 % for fallow and 5.7 % for tiger-bush as described by Allen et al. (1994).

Even if the idea that the Sahara is systematically expanding is no longer accepted wisdom (although there are many references to desert expansion in both popular media and some scientific articles), the notion of widespread land degradation as a result of human activity has persisted. Modelling studies continue to suggest that regional-scale changes in land surface properties such as albedo may significantly modify Central African climate (Xue, 1997; DeRidder, 1998). Other such studies have simulated the effects of grazing on the land surface (Stroosnijder, 1996).

There is a tendency for such studies to presuppose that land degradation is a major feature of the Central Africa environment on a regional scale, and modelling studies have tended to impose large changes in land surface properties over large areas (Xue and Shukla, 1993; Polcher and Laval, 1994; Zheng and Eltahir, 1997). Rowell and Blondin (1990) used a fully interactive model of soil wetness to investigate the impact of changes in the surface hydrology on Central African rainfall, and found a much smaller impact on the large-scale atmospheric flow than did other researchers (Sud and Fennessy, 1984; Cunnington and Rowntree, 1986) who imposed much larger initial changes over wider areas which were maintained throughout the simulations. Taylor et al. (1997) also found that the modelled influence of surface moisture flux on the evolution of the boundary layer was limited when soil moisture patterns were generated from daily rainfall estimates in a mesoscale model with a high horizontal spatial resolution. The evidence that land degradation is severe and widespread in Central Africa is limited at best.

The most widely used estimates of the extent of degradation are from UNEP (1992), based on the Global assessment of Soil Degradation (GLASOD) for the period from 1950 to 1980 (Feddema, 1998). However, this dataset has been constructed by extrapolating relatively few data, which are generally concentrated around inhabited areas, where human impacts on the land will be greatest. Wint and Bourn (1994) found livestock numbers (including the livestock associated with pastoral activity) to be highly correlated with the distribution of permanent human settlements, and only very weakly related to the extent of rangeland. This suggests that widespread overgrazing of the Central African rangelands is unlikely away from settled areas.



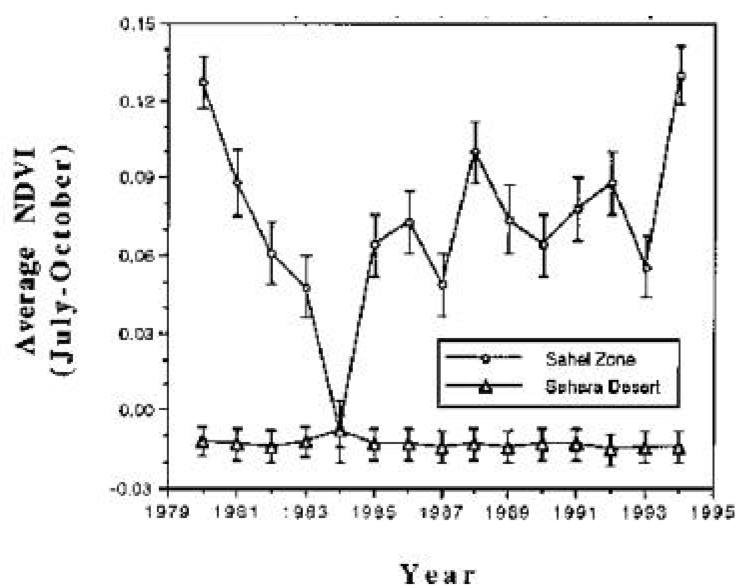


Figure 3: Relationship between NDVI and rainfall (top), and interannual variability of NDVI from 1980 to 1994 (buttom). Figures reproduced from Tucker (1995)

Prince et al. (1998) used remotely sensed vegetation indices to calculate the rainfall use efficiency (RUE) - the ratio of net primary production (NPP) to precipitation for the entire Sahel (12-18°N latitude), for the period 1982-1990. They found NPP to be in step with rainfall (reflecting the findings of Tucker et al. (1990, 1994)), with little variation in RUE, indicating resilience in the ability of the regional ecology to recover from drought that is not consistent with widespread sub continental-scale desertification. The impact of livestock on the land surface has also been investigated indirectly by Hanan et al. (1991), who found no consistent relationship between primary production (determined by NDVI values derived from NOAA AVHRR data) and proximity to wells at a resolution of 1.1 km in the North-Ferlo region of Senegal.

The assumption that traditional land use practices have a detrimental effect on the Central African environment which leads to land degradation and desertification is also questionable, even if the lack of detectable regional-scale degradation allows for localised anthropogenic vegetation clearance and soil disturbance. Many of the theories of anthropogenic land degradation are based on the idea of carrying capacity, which has been questioned as being inappropriate to agricultural and pastoral systems in Central Africa (Mortimore, 1998). Mortimore (1998, pp150-151) also discusses evidence from Kano in northern Nigeria that soils remain stable under traditional cultivation methods, even in a region with a high, and increasing, population density. Timmer et al. (1996) describe indigenous management techniques of trees in central Burkina Faso as contributing to a sustainable use of tree resources.

It appears that the susceptibility of the Central African environment to degradation and desertification, and the role of human activity in initiating or enhancing such processes, has been over-emphasised. While sand encroachment, overexploitation of natural resources and severe soil erosion may well be problems in some areas, they are not as ubiquitous as previously believed by many researchers. The type of large-magnitude regional-scale changes in the land surface that have been modelled might well have the potential to modify the Central African climate.

However, it appears increasingly unlikely that changes due to sand encroachment, overexploitation of natural resources and severe soil erosion may have taken place. Detecting any such changes is complicated by the high degree of interannual variability in the regional environment and climate. Terms such as desertification and degradation can often be misleading, and have been used to describe land surfaces exhibiting the lack of vegetation cover characteristic of dry periods or multi-year dry episodes. Much of the "detected" desertification and land degradation is likely to be associated with the temporary southward retreat of vegetation zones during the low-rainfall years that have dominated the climatological record since the late 1960s. This is not to suggest that land - atmosphere interactions are unimportant; strong wet-season feedback processes involving greater convective activity and hence greater rainfall generation over wetter Central African surfaces have been described by Taylor and Lebel (1998) and Lare and Nicholson (1994).

However, such processes are unlikely to be important on an interannual timescale, as the moisture content of the Central African soils at the onset of the wet season will be very low. Patterns of spatial persistence in rainfall due to moisture feedback are likely to be the result of the atmospheric dynamics of particular wet seasons. The lack of evidence for the kind of coherent, regional-scale changes in the Central African land surface which are associated with climatic changes in modelling studies, suggests that the origins of the Central African dry episode lie outside of the African continent.

2.3.2 The role of the oceans and global temperature patterns

2.3.2.1 Ocean and atmosphere interactions

Newell and Hsiung (1987) postulate that warmer southern hemisphere oceans and cooler northern hemisphere oceans may be the result of a decreased northward oceanic energy flux. This flux may be controlled either by the energy received at the surface of the Indian Ocean, or by changes in wind patterns in and towards the north of the tropics. Street-Perrott and Perrott (1990) calculated a mean modulus of 15 w/m^{-2} (averaged over the world ocean) for heating anomalies arising from differences in global July-September SSTs between five wet and five dry years in Central Africa. They compare this with the figure of 4 w/m^{-2} estimated to be the consequence of a doubling of atmospheric CO_2. They suggest that the north-south SST oscillation modulates the moisture convergence into the inter-tropical convergence zone (ITCZ), and has some influence on the location of the zone of strongest convective activity within it. Wagner (1996) claims that a cool North Atlantic is accompanied by an anomalous southward pressure gradient associated with weaker cross-equatorial flow from the south, strengthened north-east and slightly weakened south-east trade winds. He states that such a pattern is accompanied by increased rainfall in the Brazilian Northeast. It is likely that such a weakening of air and moisture transport across the equator from the south would also serve to reduce the intensity of the West African Monsoon, so bringing drier conditions to Central Africa.

Janicot (1994) presents evidence that since 1970 (but with the exception of 1987) ENSO events have been associated with ascent over the eastern inter-tropical Pacific and descent over the Gulf of Guinea and West Africa. This enhanced subsidence could certainly reduce the rainfall over Central Africa if the timing of the ENSO event was such that the descending motion in this region occurred during the months of June to September. A similar mechanism of convection suppression has been postulated by Shinoda and Kawamura (1994), who argue that an inter-decadal scale warming of the Indian Ocean might result in enhanced upward motion over that region, with associated descent over West Africa, thus reducing Central African rainfall. The hypothesis is interesting in light of the observed warming of the Indian Ocean and southern hemisphere oceans, and the cooling of the northern hemisphere oceans, in the 2nd half of the twentieth century (Folland et al., 1986)

2.3.2.2 The thermohaline circulation

The temperature of the North Atlantic may also be modulated by the strength of the thermohaline circulation (Manabe and Stouffer, 1988; Street-Perrott and Perrott, 1990), with cooling of the North Atlantic triggered by a reduction in the North Atlantic Deep

Water (NADW) formation. Manabe and Stouffer (1988) describe a GCM simulation with the thermohaline circulation switched off. The North Atlantic surface waters that would otherwise sink in and around the Norwegian Sea are not replaced by warmer, more saline water from low latitudes. As a result, the surface salinity and temperature of the North Atlantic Ocean are maintained at a substantially lower level than in the case with thermohaline circulation. Studies by Schiller et al. (1997), using a coupled ocean-atmosphere GCM, suggest that input of fresh water into the Labrador sea (albeit over a period of 250 years) can result in the reduction, and ultimately shutting down, of the Atlantic thermohaline circulation, reducing northward heat transport and hence SSTs in the North Atlantic. Cai et al. (1997) also describe the results of a modelled, imposed North Atlantic high-latitude freshening, equivalent to about eight times the salt deficit observed in the "Great Salinity Anomaly" of the late 1960s and early 1970s, and imposed over five model years. They report initial cooling in the sinking region, and a reduction in the NADW formation, which recovers within twenty years of the freshening being removed. However, warming occurs to the south of the sinking region and subsequently aids the recovery of the NADW formation. Other coupled ocean-atmosphere model studies and theoretical considerations suggest that the collapse of the thermohaline circulation might be a consequence of large increases in atmospheric greenhouse gases (Broecker, 1997).

Although such studies are not conclusive, the possibility that enhanced greenhouse warming may lead to a cooler and fresher North Atlantic due to the melting of Arctic ice, or freshening due to increased precipitation, should not be discounted. McCartney et al. (1996) report the spreading of unusually cold waters from the Arctic into the northeast Atlantic. Such cold temperature anomalies may be due to the input of cold water resulting from the melting of Arctic ice (Cavalieri et al., 1997) and permafrost (Hecht, 1997). Perez et al. (1995) report, from observed data, some freshening of the North Atlantic along 42°N commencing in the 1970s and persisting until 1990. Read and Gould (1992) have found that the waters between Greenland and the United Kingdom were 0.08°C and 0.15°C colder in August 1991 than in 1962 and 1981, respectively, and slightly less saline than in 1962. Melting of Arctic ice may also explain decreased salinity in the northern North Atlantic. Cavalieri et al.(1997) report that the Arctic ice cover has decreased at a rate equivalent to 2.9 ± 0.4 % per decade from 1978-1996, although this decrease has not been smooth. Four lowest summer extents before 1997 occurred after 1990. The equivalent trend in the Antarctic was of an increase in ice extent of 1.3 ± 0.2 % per decade. The authors of this study describe these results as being consistent with the modelled response to CO2 induced climate warming. Reid et al. (1998) link decreases in phytoplankton in the northeast Atlantic with the negative SST anomalies and suggest that this relationship constitutes a vegetation response to climate forcing. From 1973 to 1976 Arctic ice extent increased (Cavalieri et al., 1997), although the behaviour of the ice sheet prior to this period is unclear. Street-Perrott and Perrott (1990) describe the Great Salinity Anomaly as lasting from 1968-1982, and as being associated with deep water cooling and freshening in the Norwegian and Labrador Seas. This event could have been associated with a reduction in ice cover and consequent input of fresh water into the ocean, with anomalous atmospheric circulation or with an increase in precipitation over land areas between 35° and 70°N over the previous few decades (Bradley et al., 1987). Both this freshening and cooling episode, and another outbreak of relatively fresh water from the Arctic from around 1908-1914, coincided with drought in Central Africa. The existence of the earlier episode suggests that any apparent associations between North Atlantic freshening and enhanced greenhouse warming should be treated with caution, as the observed global and hemispheric warming has occurred since 1920 (Parker et al., 1994).

The question of when the downward trend in precipitation in Central Africa commenced is especially relevant to the hypothetical relationship between drought in Central Africa and North Atlantic freshening. If freshening and cooling of this region is important as a drought inducing mechanism the implication is that the process of desiccation should be viewed as commencing in the late 1960s, when Central African rainfall anomalies relative to the twentieth century mean were negative for the first time since 1949. If other evidence supports the view that reductions in rainfall in the Central Africa prior to 1968 should be interpreted as part of the same drying trend then changes in the Central African rainfall regime preceded the North Atlantic freshening, making the latter an unlikely candidate for a drought inducing mechanism. Nonetheless, the question is raised as to whether rainfall in the Central Africa is modulated to some extent by the behaviour of the North Atlantic Ocean, with dry conditions being associated with a freshening and consequent cooling of the North Atlantic and perhaps a reduction in the strength of the thermohaline circulation. If this is the case, it is also pertinent to ask if such processes are occurring at the present time as a response to anthropogenic greenhouse warming. If this is so, prospects for a return to generally wetter conditions in the Central Africa within the near future may be remote.

2.3.2.3 The impact of sulphate aerosols

Another potential anthropogenic modulation of sea surface temperature could arise from the impact of atmospheric sulphate aerosols produced by industrial processes. Over two decades ago, Bolin and Charlson (1976) suggested that the scattering of solar radiation by sulphate aerosols could be equivalent to a drop in the average temperature of the northern hemisphere of 0.03°C to 0.05°C. Blanchet (1995) describes fine sulphate particles as the most effective aerosol cooling agent when mixed with condensed water in moist air. Sulphate aerosol particles have a short residence time in the atmosphere (a few days), resulting in large spatial variability in their distribution (Charlson et al., 1992). As most of the world's industry is concentrated on the northern landmasses, atmospheric sulphate aerosols are generally confined to the northern hemisphere. In modelling studies conducted by Haywood et al. (1997), the radiative forcing due to sulphate aerosols in the northern hemisphere was a factor of four greater than in the southern hemisphere, although forcing was strongest over and downwind of sources, i.e. over land areas.

An earlier assessment of the empirical evidence by Charlson et al. (1992) suggested that this climate forcing was twice as great in the northern hemisphere as was the global average. These authors also state that aerosol forcing is greatest in the daytime in summer, and that concentrations of such particles are an order of magnitude greater within 1000 km of the sources than in remote regions. Charlson et al. (1992) report that tropospheric mass concentrations of non-sea-salt aerosols sulphate over the northern hemisphere oceans appear to be enhanced by 30% or more over the natural background. Erickson et al. (1995) hypothesise that sulphate aerosol-induced changes in cloud albedo may result in cooling of the North Atlantic and North Pacific, with the largest effects occurring in winter. This suggestion is the result of modelling studies, in which circulation changes result from the different responses of the North American and Eurasian landmasses, with a deepening of the trough over the former, and an enhanced ridge over the latter. Inclusion of sulphate aerosols in climate models has resulted in simulations reproducing the observed record more accurately due to the sulphate aerosol cooling effect (Mitchell et al., 1995a, b). However, the radiative processes involved are poorly understood and it is possible that the aerosol representation in the simulations is not very realistic (Cane et al., 1997). Inclusion of aerosol representation alongside greenhouse gases (GHGs) in a simulation using the Hadley Centre climate model led to closer agreement of the modelled and observed global changes in temperature, particularly in the most recent decades, when sulphate aerosol forcing has been largest (Mitchell et al., 1995). In this simulation the largest decrease in temperature, relative to the GHG-only case, due to the aerosol forcing occurred in the northern mid-latitudes and the Arctic. The former regions are subject to the greatest aerosol forcing, while cooling in the Arctic is amplified by increases in sea ice. A model study of the effects of sulphate aerosols alone by Taylor and Penner (1994) resulted in the largest cooling occurring not in the regions of greatest forcing, but over the Norwegian and Greenland seas. Again, this cooling in the simulation was associated with increases in sea ice. It should be noted that this particular modelled response of the climate system to increased sulphate aerosol loadings conflicts with the observations presented by Cavalieri et al. (1997). They describe decreases in the extent of the Arctic sea ice from the late 1970s, the period during which Mitchell et al. (1995b) suggest the effects of sulphate aerosols (at least in terms of radiative forcing) should have been greatest. Hulme (1998) found that modelling studies performed at the Hadley Centre (the HadCM2 simulations) incorporating the effects of sulphate aerosols did not simulate the Central African dry episode. These studies therefore do not support the hypothesis that such aerosols reinforce the temperature anomaly dipole associated with a dry Central Africa. However, it is not clear that the simulations capture the relationship between rainfall and global temperature patterns.

2.3.2.4 Central African rainfall and twentieth century global and hemispheric air temperatures

It is interesting to compare the above SST-related studies with those of global surface air temperatures whose principal goal is the detection of global warming signals. Jones and Briffa (1992) state that the warming which is apparent on a global scale over the course of the twentieth century occurs mostly in winter and spring and exhibits a notable spatial variability in the northern hemisphere, whereas in the southern hemisphere warming is evenly distributed throughout the year and generally more spatially homogeneous. Annual warming in the twentieth century is greatest in the southern hemisphere, and cooling is apparent over the northern Pacific Ocean and northern Atlantic Ocean for the decade 1981-1990, which contained several highly rainfall-deficient years. It may be speculated that the reduction in Central African rainfall is the result of a global warming which has resulted in a preferential heating of the southern hemisphere, due in part to the greater negative solar forcing due to aerosols in the northern hemisphere (Roeckner et al., 1995).Variations in global and hemispheric temperatures over the twentieth century have also been described by Parker et al. (1995). Figure 4 shows the series of annual average global and hemispheric combined land-air and sea-surface temperature anomalies of Jones (1994); Parker et. al. (1995).

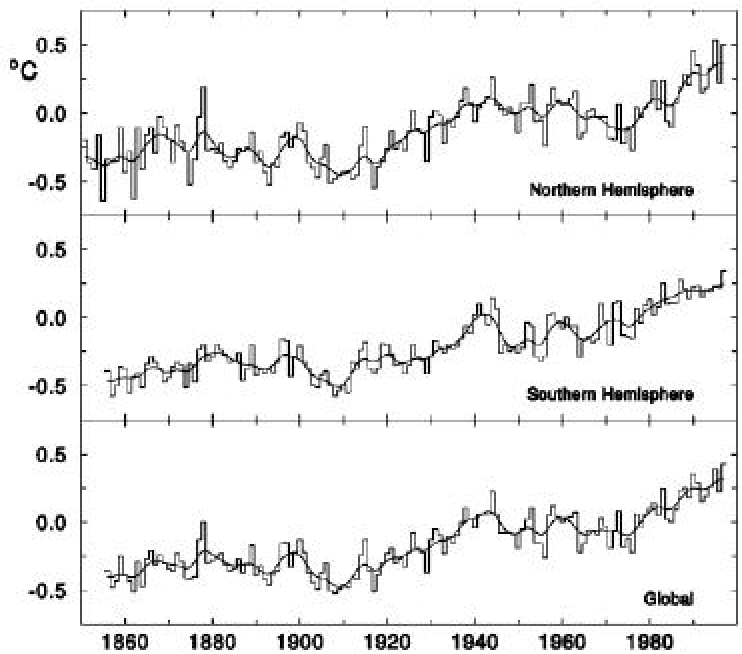


Figure 4**:** Combined annual land air and sea surface temperature anomalies relative to 1961-1990, based on the datasets of Jones (1994) and Parker et al. (1995)

A global and hemispheric warming commences around 1920, after the dry year of 1913 in Central Africa. The increase in temperature ceases in the 1940s, and in the early to mid 1940s there is a slight cooling of both hemispheres. The shape of the anomaly curves is similar for both hemispheres until the early to mid 1960s, exhibiting no general trend, and with the southern hemisphere exhibiting mostly negative anomalies and the northern hemisphere mostly positive anomalies. Parker et al. (1994) describe a slight fall in global average temperatures, with a relatively warmer northern hemisphere in the decade 1951-1960, when Central Africa experienced abundant rainfall. After 1965 the southern hemisphere warms, but the northern hemisphere anomalies do not increase until the late 1970s. The decade 1971-80 was characterized by a relatively warmer southern hemisphere, and coinciding with the Central African drought of the early 1970s. So a change in the inter-hemispheric contrast (IHTC) to the pattern associated with drought in Central Africa appears to have occurred in the decade 1961-70. This again suggests that the downward trend in rainfall should be viewed as commencing in the late 1960s, and that the reduction in rainfall which is apparent from the mid-1950s represents merely a return to “normal” conditions after a wet episode related to a particular mode of SST variation.

The northern hemisphere series exhibits negative anomalies in 1984 and 1985, with southern hemisphere anomalies remaining positive. The two largest rainfall deficits in Central Africa were in 1983 and 1984. Parker et al. (1994) state that during the most recent warming period (i.e. since 1980) much of the central and western North Atlantic north of 50°N and much of the central and mid-latitude North Pacific have cooled. It is plausible that the cooling of these regions is related to changes in ice cover and freshwater input, as has been observed in modelling studies (Mitchell et al., 1995a). An important question concerning the association between Central Africa rainfall and the IHTC is whether rainfall is directly modulated by circulation changes which are driven by the SST patterns or whether both Central African rainfall variations and changes in the IHTC are manifestations of more fundamental climate processes.

Early theories concerning the causes of drought in Central Africa concentrated on the mean position of the ITCZ, and the northward extent of its migration in the northern hemisphere summer. Lamb (1977) suggested that drier conditions in the region were the result of changing patterns of atmospheric pressure in mid- to high-latitudes, with an expanded region of cyclonic activity in the former resulting in the displacement of the tropical climatic zones towards the equator by a few degrees. Lamb (1978) describes the near-equatorial pressure trough as being some 200 km south of its mean position from July to September in dry years, and anomalously deep, with an associated anomalous southward extension of the North Atlantic high-pressure region. He states that this pattern is not seen in earlier months, although such periods exhibit a stronger than normal horizontal pressure gradient, separating higher than normal pressure north of 20°N from lower than normal pressure south of 12°N, with a reduced northward extension of the South Atlantic high pressure region. Citeau et al. (1994) suggest that the strengthening of the South Atlantic anticyclone and the associated enhanced subsidence extend further east thus producing stronger southeasterly trade winds which will strengthen the West African Monsoon.

During periods exhibiting the cold northern hemisphere and warm southern hemisphere SST anomaly pattern, the equatorial Atlantic experiences stronger northerly winds to the north of the equator, and weaker southerly winds to the south of the equator. The latter weaken the upwelling of cold water off the African coast (Nobre and Shukla, 1996). The largest wind anomalies occur between 10° and 20°N, and are associated with strong sea-level pressure (SLP) anomalies further north. Nobre and Shukla (1996) suggest that during the period March to May, wind stress anomalies are a response to the meridional SST gradient, which is strongest in this period. The SST anomaly pattern moves westwards from the December to February to the March to May period, with the leading eigenvector of the meridional wind, whose amplitude is at a maximum in this season, over the western tropical Atlantic, being directed towards the warmer waters of the southern hemisphere. Thus the meridional component of the wind is determined by the meridional SST gradient, but the subsequent development of the SST anomalies is a response to variations in the strength of the trade winds.

Shinoda (1990) attributes poor rains in Central Africa primarily to a reduction in the total rainfall of the monsoonal rainbelt rather than displacement of the belt towards the equator, although such a southward displacement of the centre of gravity of the rainbelt is observed during August in dry years. An analysis of atmospheric water vapour content (AWVC) at Bidi (northern Burkina Faso at 15º50’N) over the period January 1987 to September 1989, and over the 1985/86 and 1986/87 dry seasons at Ouangofitini

(Côte d'Ivoire at 9º36'N) by Faizoun et al. (1994) found good agreement between the AWVC values for these periods and climatological values obtained before the onset of the Central African dry episode. They infer from this that there is no significant link between monthly AWVC values and the low rainfall values experienced in the region since the onset of desiccation. This is a significant result, although the extrapolation of such a conclusion should be treated with caution as long as the data are restricted to only three wet seasons. The 1987 wet season was characterized by a large rainfall deficit (approximately 30% relative to the twentieth century mean). Rainfall in 1988 was close to the mean for the twentieth century. 1989 was characterized by a rainfall deficit of intermediate magnitude, some 10% below the mean. Conclusive evidence that rainfall and AWVC are not generally related would require comparisons of these two quantities for many years of varying rainfall, including several years with large rainfall deficits. Nonetheless, these findings beg the question as to whether drought conditions arise from a lack of convective activity rather than from lack of precipitable moisture. Such a hypothesis is consistent with the conclusions of Bell and Lamb (1994), who have found that the decline in the monsoonal rainfall in the Central Africa results from a general progressive decrease in the size and intensity of the disturbance lines (DLs) which generate most of the rainfall over the Central Africa. These are organized lines of westward propagating convective cells arising from baroclinic and barotropic instabilities in the African Easterly Jet whose generation and development also depend on orography, the availability of surface moisture and surface temperatures (Peters and Tetzlaff, 1988; Rowell and Milford, 1993). Shinoda (1990) presents evidence that the recent dry episode in Central Africa is the result of a weakening of the African Hadley circulation, and that continuous intensification of the northern subtropical high has contributed to this weakening. He describes increases in 700 hPa heights and temperatures throughout the zone 30°N-30°S in August, about east-west axes over Central Africa, and attributes such increases to enhanced subsidence heating.

Other studies have also linked dry conditions in the Central Africa with a weakened northern hemisphere circulation. Eltahir and Gong (1995) compare the years 1958 and 1960, and find weaker August-September meridional circulation, particularly with respect to southerly flow, in the drier year of 1960. By contrast, the meridional circulation in 1958 is strong, characterised by southerly winds from the surface to 300 hPa, and northerly winds from 300 hPa to 100 hPa. The August-September zonal wind at the equator is stronger in 1958. Landsberg (1975) described rainfall over Central Africa as being essentially determined by the strength of the southern Hadley cell. Thus a reduction in Central African rainfall is associated with a weakening of the northern hemisphere Hadley circulation, which has been synchronous with a large scale tropospheric warming throughout the northern hemisphere. Such a trend is consistent with the expected enhanced greenhouse warming resulting from anthropogenic emissions of greenhouse gases (IPCC, 1996). It is possible, therefore, that a spatially heterogeneous anthropogenic warming of the troposphere has modified the atmospheric circulation in such a fashion as to reduce the strength of the West African Monsoon. However, such theories are highly speculative at the present time.

2.4 The importance of “internal” and “external” processes

Central Africa has been subject to a change in the nature of the rainfall regime which is determined by the behaviour of the West African Monsoon. This has resulted in a desiccation of the region since the 1970s and higher interannual rainfall persistence than prevailed in the earlier part of the twentieth century. The desiccation appears to be largely synchronous with large-scale changes in atmospheric circulation, characterized by a general warming of the troposphere and a reduction in the intensity of the northern hemisphere meridional circulation. Also associated with these trends is the establishment in recent decades of a north-south dipolar sea surface temperature pattern, with anomalies in the northern and southern hemispheres generally being out of phase (but with the whole Indian Ocean exhibiting the same behaviour as the southern hemisphere oceans). It is therefore reasonable to say that the recent Central African desiccation is associated with large-scale changes in the global coupled ocean-atmosphere system.

Dry conditions in tropical North Africa are not necessarily synonymous with a shallower monsoon layer, or with a reduced northerly penetration of the Inter-Tropical Convergence Zone, but are associated in a more general fashion with the intensity of the monsoonal airflow. Weakening of the monsoon may be viewed as being due to the general weakening of the Hadley circulation on a hemispheric or global scale. Lag relationships between temperatures in the Atlantic and Pacific Oceans, and the correlation of Central African drought with particular large-scale patterns of SST, support the notion that conditions in Central Africa are a response to global scale processes. However, the relationship between Central African rainfall and hemispheric oscillations in SST anomalies is not simple, and does not hold on an interannual time-scale over the entire period of desiccation, implying that other mechanisms also affect rainfall in the region. Such mechanisms could be the result of atmospheric teleconnections which are relatively independent of the mean large-scale circulation, or could result from local and regional scale feedback processes. If the breakdown of the rainfall-IHTC relationship on short time-scales is the result of persistence in the regional climate system, then feedback processes operating on a synoptic scale over tropical North Africa appear to offer the most likely explanation for such discrepancies.

One of the principal factors in determining the strength of the West African Monsoon will be the distribution of regions of convergence and divergence throughout the troposphere, and the intensity of the resulting convection and subsidence. For example, variations in the strength and extent of the sub-tropical highs will impact on monsoonal airflow. The strength and extent of subsidence over North and West Africa will also be important, and it is highly plausible that regional factors such as land-atmosphere interaction may modulate atmospheric stability. Modelling studies of changes in soil moisture and vegetation cover were discussed briefly in Section 2.4, and doubts were raised about the degree to which such studies accurately represent changes in these quantities. It was seen that it is highly likely that the modelling studies have greatly exaggerated the extent and severity of changes in the land surface. One land-atmosphere interaction which has not been widely studied, particularly from a modelling perspective, is the mobilisation and transport of atmospheric dust and its subsequent effect on the radiative and dynamical properties of the atmosphere. The influence of global SST patterns and changes in the global or hemispheric atmospheric circulation in this sense can be viewed as "external" to the region, or not necessarily dependent on conditions within the African continent. Of course "internal" and "external" processes will interact, with the latter exerting a greater influence on the former than vice versa (although a dust-induced cooling of the tropical North Atlantic as postulated in Section 2.5.5 provides a possible mechanism by which internal processes could influence the external forcings).

A conceptual model can be described in which large-scale changes in the behaviour of the atmosphere and oceans precipitate drought in Central Africa. The increased aridity in Central Africa will result in changes in the local and regional land-atmosphere system, which may act to reinforce and sustain drought conditions even after the removal of the external factors which initiated drought. In order for a return to wetter conditions to occur, atmospheric or oceanic anomalies of the similar magnitude and opposite sign to the original drought-inducing anomalies must occur. Such a scenario has been postulated by Nicholson (1995).

3 Data and Methodology

In this chapter relevant data are introduced and tested for homogeneity. The daily station data derived from weather stations across Central Africa and gridded temperature and precipitation dataseries reanalysed by NCEP/NCAR (Kalnay et al., 1996) are described in section 3.1.2 and 3.1.3 respectively. Mathematical methods applied to climate data are detailed in Section 3.3.

3.1 Climate data

Data must be available in order to accurately determine meaningful climatic trends, temporal and spatial patterns over Central Africa for periods of decades (Karl et al., 1999; Nicholls and Murray, 1999; Frich et al., 2002). It is also important that each series under consideration possesses a complete or near complete set of high quality values (Manton et al., 2001). This causes problems when data from Central Africa, much of which has suffered civil unrest during recent years are analysed. Africa's weather monitoring system is deteriorating and needs major improvements to meet the challenges of climate change determination. Overall it is estimated that Africa needs 200 additional automatic weather stations, a major effort to rescue historical data and improved training and capacity building on climate and weather reporting. Decades ago, Africa had a relatively dense network of stations to measure rainfall temperature and other weather data but now many these stations are silent. Figure 5 shows a map of the locations of weather stations on the globe and the percentage of activities at those stations, the figure reveals a significant amount of silent stations in Africa. About one in four weather stations in eastern and southern Africa that are part of the international data-sharing network called the Global Climate Observing System are not working, and most of the remaining stations are not functioning properly. This acts as a huge brake for climate science, the data scarcity hampers climate researchers around the globe, who need historical data and input from current events to improve their models. The reasons for the data scarcity include a degree of protectiveness; some centres in Africa fear that releasing data might allow outside companies to generate forecasts for agriculture, aviation and other industries; a function currently performed by state-run agencies. Others argue that historical records are viewed as bargaining chips. It may not be a bargaining tool; it may be because centres can't make the data available as some measurements, such as those from balloons, are often too expensive (Christen, 2004).

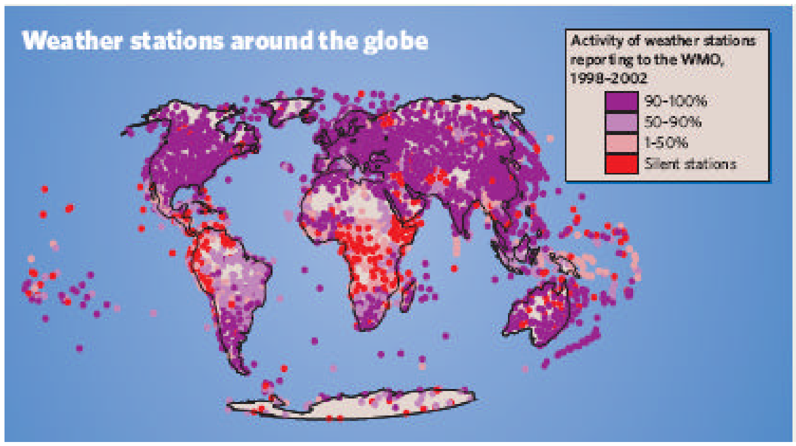


Figure 5: Locations and activities of weather stations reporting to the WMO 1998-2002; source: WMO/UK Department for international Development

The lack of station activity is highlighted in two station datasets from the National Oceanic and Atmospheric Administration and the Climate Research Unit obtained for the purpose of this study. The number of meteorological stations supposed to be measuring data was 2307 in both datasets; however an analysis to determine data from each station within the datasets revealed that only a few of the stations actually recorded data. For example in the NOAA station dataset, only 20 stations actually recorded data within the study area and of the recorded data 65.5% of the data was missing (refer to Table 1), meaning critical climate information which may be important for climate research has been lost and may render data from such stations unsuitable for long term trend analysis. Only three i.e. Ada (65475, DGAD), Faya (64753, FTTY) and Accra (65472 DGAA) recorded enough data (26.6, 23.97 and 33.94% missing data in temperature data set). This is not sufficient to present results reflective of the climatic conditions prevailing in the whole study area but could provide important information as a check for results obtained with other datasets.

Table 1: Percentage of missing data in NOAA dataset for a period of observations from 1987-2003.

	Temperature (ºC)		***Precipitation*** (mm)	
Station name	***No. of missing values***	***% of missing values***	***No. of missing values***	***% of missing values***
Accra (65472,DGAA)	1982	33.94%	2543	43.54%
Ada (65475,DGAD)	1557	26.66%	1700	29.11%
Akatsi (65462)	3820	65.41%	3934	67.36%
Abong-Mbang (64960,FKAG)	4097	70.15%	4215	72.17%
Betare-Oya (64901, FKAO)	3662	62.71%	4433	75.91%
Garoua (64860,FKKR)	3096	53.01%	3437	58.85%
Bousso (64705)	5154	88.25%	5215	89.30%
Fada (64757)	5009	85.77%	5029	86.11%
Faya (64753, FTTY)	1400	23.97%	1391	23.82%
Bunia (64076)	3616	61.92%	4029	68.99%
Bunia -Ruampara (64077)	4380	75.00%	4531	77.59%
Buta (64034)	4488	76.85%	4614	79.01%
Kano (65046, DNKN)	4535	77.65%	4681	80.15%
Katsina (65028)	4400	75.34%	4591	78.61%
Lagos/ Ikeja (65201, DNMM)	3686	63.12%	3759	64.37%
Lagos/Oshodi (65202	5507	94.30%	5537	94.81%
Makurdi (65271, DNMK)	2789	47.76%	3347	57.31%

3.1.1 Station data

3.1.1.1 National Oceanic and Atmospheric Administration (NOAA) station data

The global station dataset from NOAA consists of measured station data from all meteorological stations in Africa and can be obtained from the African Desk of the NOAA Climate Prediction Centre. The dataset includes temperature and precipitation data measured in degrees Celsius and millimetres respectively. The temperature data are measurements of minimum and maximum daily temperature and the precipitation data are measurements of 24-hour precipitation. As stated above data from over 20 stations across Central Africa were obtained from the global station data set of the National Oceanic and Atmospheric Administration.

All data from 1987-2002 is provided, except for December 1988 which was corrupt and unavailable. The data is zipped in yearly files each containing 12 files with monthly data. The monthly files are of the format = nwdlyYYYYMM.dat where the YYYY represents the year and MM represents the month and for each monthly file. The data of interest reside in the first 4 columns e.g. 2001052301234 11.7 5.4 3.3

The first column actually represents two parameters:

20010523 = Date in YYYYMMDD format

01234 = ground station ID

Column 2 is maximum temperature in degrees Celsius

Column 3 is minimum temperature in degrees Celsius

Column 4 is 24-hour accumulated precipitation in millimetres

3.1.1.2 Climatic Research Unit (CRU) observational data

Another data source that can be used for climate research in Central Africa is the observational data from the Climate Research Unit (CRU) which holds global datasets of station climatological normals and monthly time series of precipitation and mean, maximum and minimum temperature (New et al., 1997). Rainfall and temperature data have been created by interpolating meteorological data onto a 1º latitude x 1º longitude geographical grid (rainfall) and 0.5 º latitude x 0.5º longitude geographical grid (temperature). The datasets are described by New et al. (1999a, b).

Data for Africa can be extracted from the master datasets. The resultant African subsets comprise of 2307 precipitation stations (Figure 6), 1485 mean temperature stations (Figure 7) and 1431 diurnal temperature range stations (Figure 8). The datasets of station time series are considerably smaller than the network of station normals and also show large variation over time. The late 1980s and 1990s are particularly poorly sampled; in some cases this is due to political and/or economic factors.

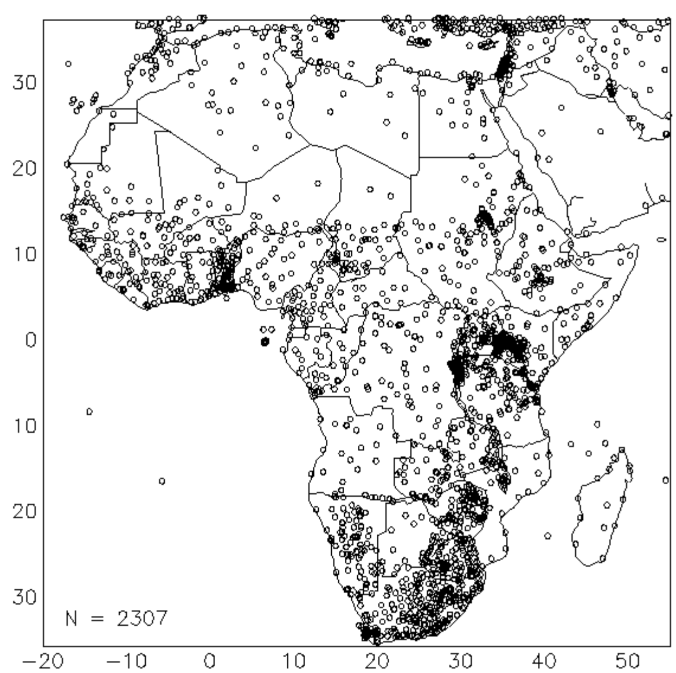


Figure 6: Station network of precipitation, New et al. (1997)

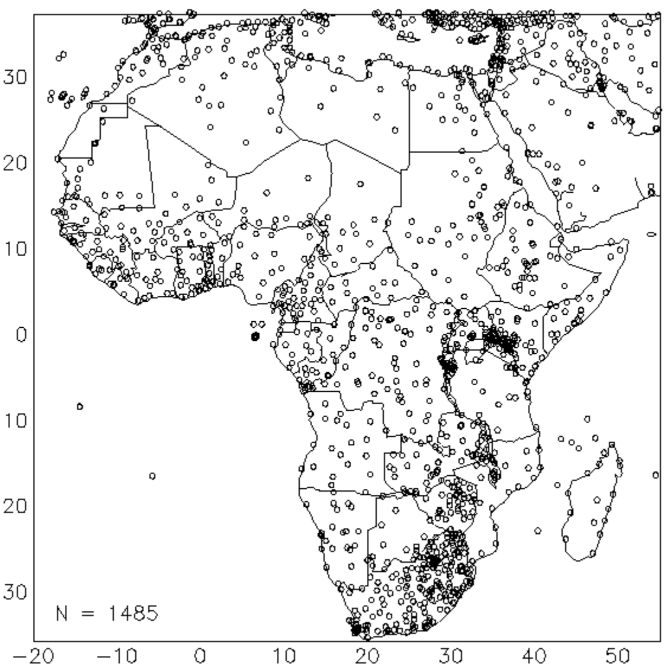


Figure 7: Station network of mean temperature normals, New et al. (1997)

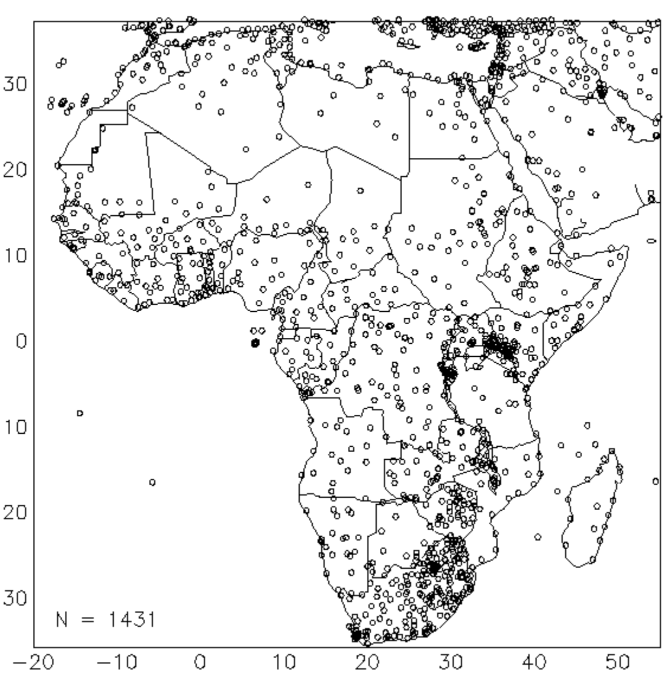


Figure 8: Station network of diurnal temperature normals, New et al. (1997)

3.1.2 Gridded reanalysis data from NCEP

The term 'reanalysis' refers to a methodology that assimilates data from a large number of sources (e.g. land, ship, rawinsonde, aircraft, and satellite observations) into a single consistent model through time (even though data availability may change with time). At NCEP/NCAR incoming data undergoes four distinct stages to test for reliability (Kalnay et al., 1996). Preprocessing compares input with pre-existing climatologies via tendency checks, homogeneity testing, box averages, and variance analysis. Parallel processing tests the assimilation of a new data source by processing a single year of data with and without the data source. The third stage is internal validation and is the most complex part of the validation process. Tests vary with data type but are often interpolative in nature. Through a combination of horizontal, vertical, hydrostatic and increment interpolation checks, a baseline elevation check and a temporal interpolation check, forecast and mean variables (such as sea level pressure, wind speed, and temperature), and errors in station locations, along with possible changes of location, may be obtained. This kind of quality control is known as 'optimal interpolation' (Woolen, 1991; Reid et al, 2001). Finally the reanalysis system undergoes rigorous (but largely automated) monitoring. In its entirety, the process is computationally intensive, and takes time for input data to be gathered and collated. Output is thus retrospective. The end result of the reanalysis process is consistent and gridded data (NCEP/NCAR has currently a 2.5 by 2.5 degree resolution) of over 50 years, at multiple heights, for a very large number of variables (Kalnay et al., 1996).

The NCEP/NCAR data were chosen over alternatives because at the beginning of this study they were unique in terms of coverage, both spatially and temporally. A number of fairly substantial errors in the dataset have been highlighted by Kistler et al. (2001) but pressure and humidity data remain some of the most robust data, rating an A (strongly influenced by observed data, e.g. upper air temperature, and wind) or a B (strongly influenced by observed data and the reanalysis model, e.g. humidity, and surface temperature) rather than the C (derived entirely from model fields, e.g. clouds, and precipitation) rating applied to the majority of surface fluxes, not used in this study (Kalnay et al., 1996). Over Africa the reanalysis data set is of 'research quality' in terms of both interannual and (less so) decadal variability, and is suitable for the purposes of this study, as for others (Quadrelli et al., 2001; Bordi and Sutera, 2002; Goodess and Jones, 2002). Even though the spatial distribution of 2.5 degree may be large, the datasets contain enough grid points to provide a representative coverage of Central Africa. For these reasons the reanalysis data was selected as the primary data source for analyses in this research.

The area of investigation has its upper left corner at 18 º N 18 º W and its lower right corner 0.75 º S 44 º E. The data were made up of 224 data files each representing grid averages on a 2.5 º lat-long box.

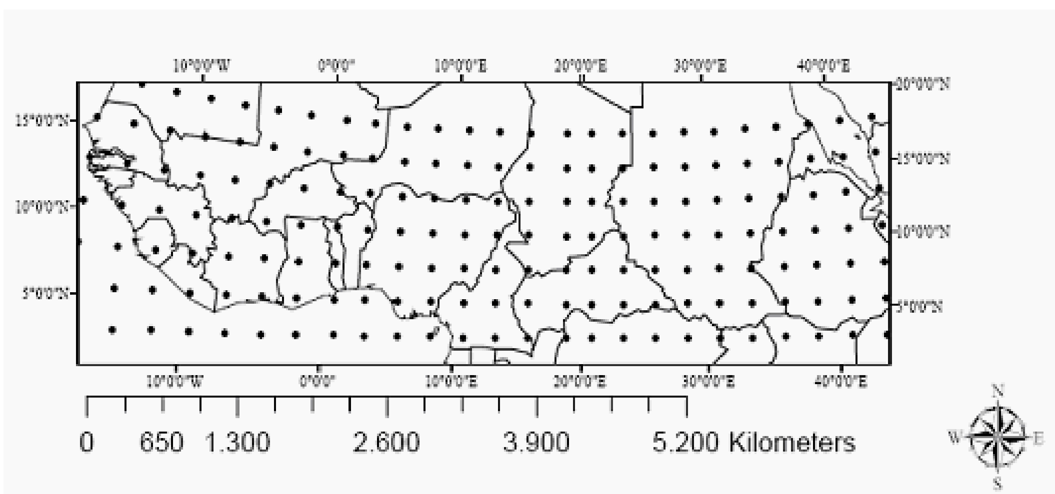
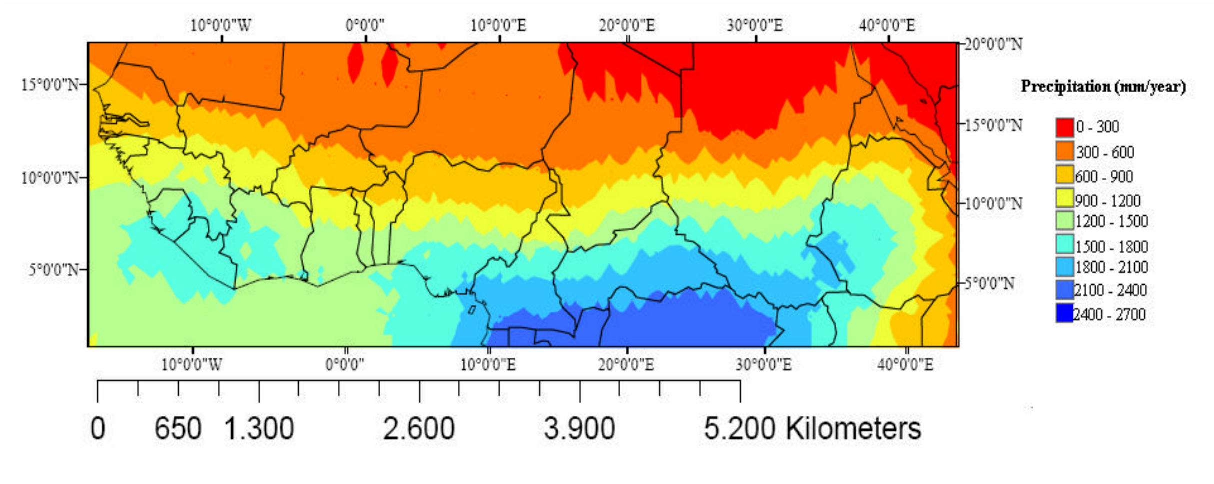


Figure 9: Map of study area with locations of grid- points separated by a 2.5 º lat-long box used to perform the analysis.

The time series and the spatial distribution of the annual averaged temperature and precipitation (1948-2004) are illustrated in Figures 10 and 11, respectively. The time series shows an increase in area averaged temperature and decrease in precipitation. Trends from these series are analysed in detail in chapter 4. The factors that may cause the distribution of the spatial patterns as illustrated in the figures include effects of land-sea distribution and mountains. Even at the same latitude, climate is different between over oceans and over land, or between coastal and interior areas. The cause of the difference is primarily the difference of heat content between ocean and land, and next the difference of capability to supply water. In addition, the fact that the land surface in Central Africa is not usually flat but has mountains affects climate.

The effect of land-sea distribution related to precipitation is that the availability of water for evaporation from the surface is indefinite over the ocean but limited over the land. It consists of two factors. (1) Even the same amount of energy may be supplied; there is not enough water for evaporation over arid lands such as deserts. (2) Because of the difference of heat capacity, the ocean can store energy in summer and evaporate water in winter using it. Evaporation over the land is determined by the energy supply within the season so that it must be small in winter. Large bodies of water have a moderating effect on temperature as well. That is, if it is cold inland, the land near the large body of water will not be as cold as it is inland. If it is hot inland, the land near the water will not be as hot. So the climate is usually milder near the water. Inland, however, there may be great differences between the coldest and warmest temperatures

The effect of mountains on the large-scale climate can be sorted into the mechanical effect concerning the motion of the atmosphere and the effect to the water cycle. (1) The motion of air is deformed by the mountains as obstacles. Its effect propagates downwinds and forms troughs and ridges of pressure. Even the position of the mountain is the same; the pattern of effects is different between the case where the flow go over the mountain and where the flow goes around the mountain. (2) As the effect to watercycle, air is forced to go up at the mountains, then water vapor condenses and precipitates near the mountain, and air depleted of water vapor is supplied to the downwind area.



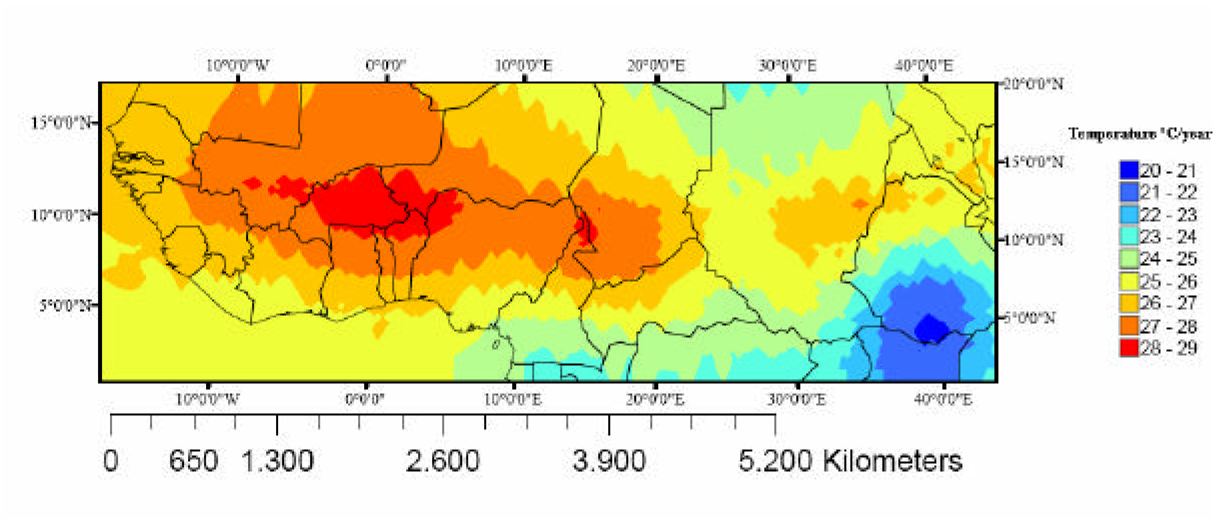
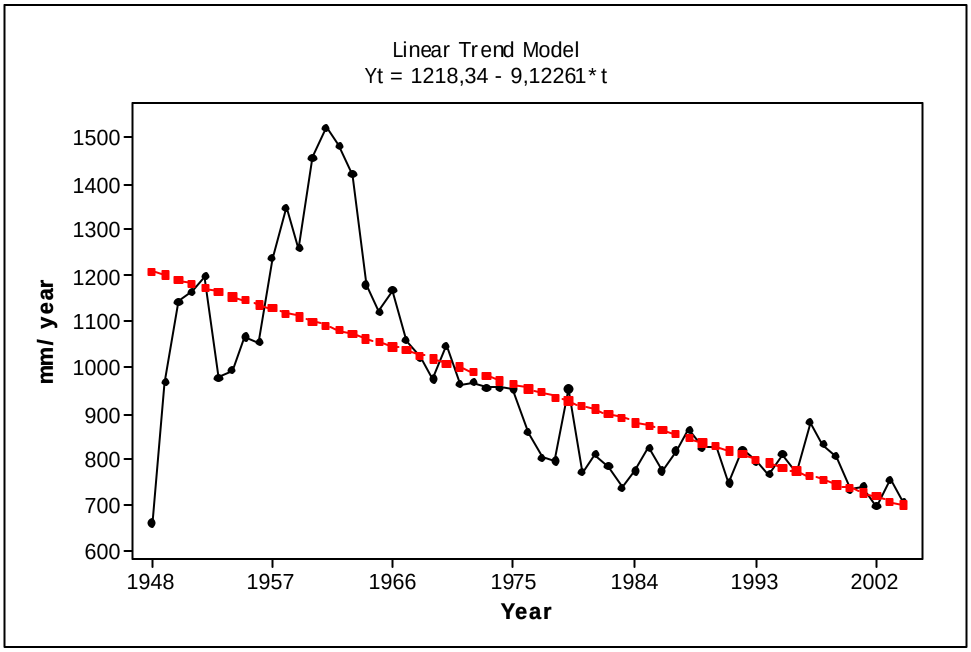


Figure 10: Spatial distribution of mean annual precipitation (top) and temperature (bottom) (1948-2004) for Central Africa as in the NCEP reanalysis data



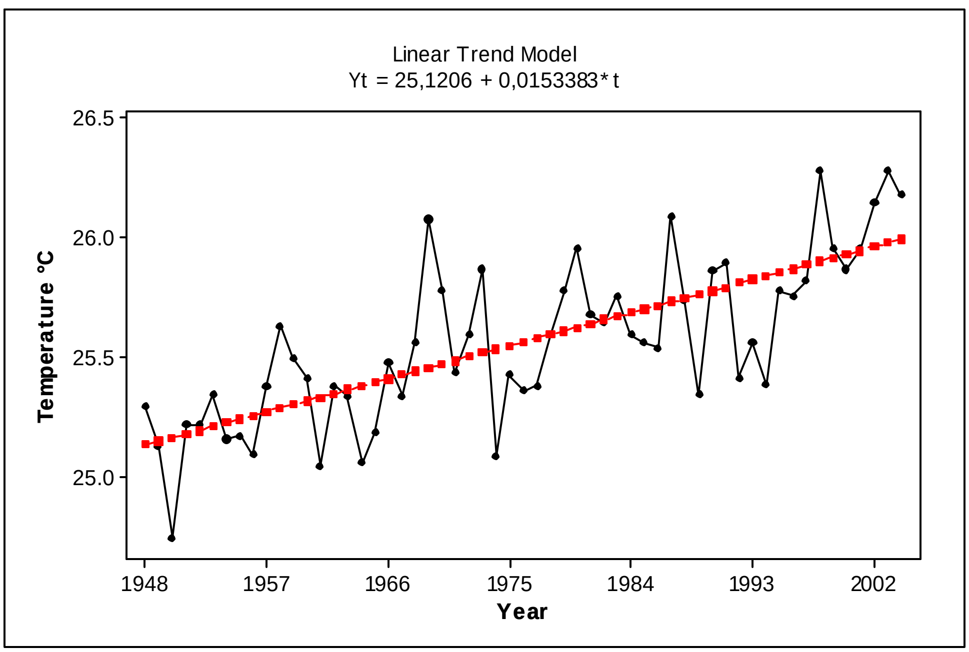


Figure 11: Temporal timeseries of mean annual precipitation (top) and temperature (bottom) (1948-2004) for Central Africa as in the NCEP reanalysis data

3.2 Homogeneity testing

Climate data can provide a great deal of information about the atmospheric environment that impacts almost all aspects of human endeavor. For example, these data have been used to determine where to build homes by calculating the return periods of large floods, whether the length of the frost-free growing season in a region is increasing or decreasing, and the potential variability in demand for heating fuels. However, for these and other long-term climate analyses –particularly climate change analyses– to be accurate, the climate data used must be as homogeneous as possible. A homogeneous climate time series is defined as one where variations are caused only by variations in climate.

Unfortunately, most long-term climatological time series have been affected by a number of non-climatic factors that make these data unrepresentative of the actual climate variation occurring over time. These factors include changes in instruments, observing practices, station locations, formulae used to calculate means, and station environment. Some changes cause sharp discontinuities while other changes, particularly change in the environment around the station, can cause gradual biases in the data. All of these inhomogeneities can bias a time series and lead to misinterpretations of the studied climate. It is important, therefore, to remove the inhomogeneities or at least determine the possible error they may cause.

Many researchers have put a great deal of effort into developing ways to identify non-climatic inhomogeneities and then adjust the data to compensate for the biases these inhomogeneities produce. Several techniques have been developed to address a variety of factors that impact climate data homogenization such as the type of element (temperature versus precipitation), spatial and temporal variability depending on the part of the world where the stations are located, length and completeness of the data, availability of metadata, and station density.

Although the best way to ensure homogeneity is to keep the record homogeneous through appropriate management of the observations site and associated equipment, this is very difficult to achieve. Besides, because it is almost impossible to be 100% sure about the quality of past data, a homogeneity assessment is always recommended. There is not one single best technique to be recommended. However, the four steps listed below are commonly followed:

1) Metadata Analysis and Quality Control.
2) Creation of a reference time series.
3) Breakpoint detection.
4) Data adjustment.

3.2.1 Metadata and quality control

Metadata are important for identifying discontinuities in a time series. By putting together all the available metadata and building the station history, we anticipate and preview what problems we may find in the data and when they should appear. Some homogenization approaches only accept discontinuities registered in the metadata. This is indeed a good approach if we believe that our metadata are absolutely complete, from the first to the last observation. When trying to detect inhomogeneities we are looking for the fingerprints of factors other than climate and weather in data. That means there is always a cause for any inhomogeneity. Should metadata be perfect, we could always identify this cause and there would be no need to employ any statistics to find further breakpoints in a time series. Nevertheless, even in the presence of the most carefully documented metadata, it is advisable to compare what the station history says and what data analysis identifies, as a sort of double check.

Another way we may benefit from metadata is to know what kind of quality control (QC) the data has undergone. QC procedures vary from very simple techniques, such as plotting the data against time (alone or together with neighboring stations) or identifying data outlying pre-fixed thresholds, to sophisticated analysis that cross validates different meteorological elements at the same station and/or data from different stations. Even in those cases when we are aware that a complete QC has been applied, it is recommended to plot the data before actually starting the homogenization procedures and correct or remove from forthcoming steps obviously wrong data. This is crucial because many homogenization techniques rely on comparing the central value from two different data sections. Failing to remove outlying data enormously complicates statistical detection of any inhomogeneity or, in the best case, alters the value of the correction factor, especially if we are using a parametric test. By plotting the data we can also identify other values, like special codes, that must not enter our analysis. For example, it is very common to use figures like "-999" to identify missing data. Failing to cut this code from the analysis will completely ruin it. Plotting the data will help us to decide if the data are free from these kinds of problems and we can go ahead with homogenization or we need to go back and further quality control them. At the same time, data plots can make us aware of obvious inhomogeneities in data.

3.2.2 Building a reference time series

Detecting and adjusting inhomogeneities is a hard and difficult task, as on most occasions the magnitude of the inhomogeneities is the same or even smaller than that of true climate-related variations. For this reason, it is advisable to create a reference time series and compare with the station to be homogenized, the so-called candidate station series. A reference time series ideally has ideally to have experienced all of the broad climatic influences of the candidate, but none of its artificial biases. Should the candidate have no inhomogeneities, when the candidate and reference series are compared by differencing (in the case of variables measured on an interval scale, like temperature) or by calculating ratios (for variables measured on a proportional scale, like precipitation), the resulting time series will show neither sudden changes, nor trends, but will oscillate around a constant value. However, if there are one or more inhomogeneities, the difference or ratio time series, will reveal their fingerprint.

The most common approach for building a reference time series is to calculate for each year a weighted average of data from neighboring stations or sections of neighboring station time series that metadata indicate are homogeneous. Some measure of similarity (usually correlation coefficient) is employed to select the most adequate neighbors and weight them according to their statistical resemblance to the candidate. Several widely used techniques calculate the correlation coefficients between first-differenced time series. A first difference series is made by subtracting year 1's observation from year 2, year 2 from year 3, etc. The correlation then is a measure of the similarity in year to year changes, and an inhomogeneity only impacts one observation rather than making all observations after the inhomogeneity artificially warmer or colder. A widely used alternative to integrate different series into an averaged reference is to compare them one by one with the candidate.

Other approaches extract principal components from the whole data network, or use an independent data source thought to be homogeneous. Creating and using reference time series may encounter two major problems, the first one being the lack of data to build them. It is also true that the difficulties for building a reference time series increase exponentially with the increase in spatial variability of the data. Spatial variability depends on three key factors: the meteorological element we are dealing with; the type of climate; and the time resolution.

3.2.3 Breakpoint identification

The fourth step is to search for breakpoints in the difference or ratio between the candidate and the reference time series (or alternatively in the data when a suitable reference cannot be built), compare them with the available metadata and decide which discontinuities will be indeed regarded as true inhomogeneities. Some methods do not actually search for breakpoints and only use the reference time series to decide if the changes found in the station history produce an effect in the data large enough to require adjustment. This is a good approach only if the metadata are believed to be complete and up to date. Common statistical tests for samples comparison, like the *t-test* or rank-based alternatives (if data normality is in doubt), are adequate to decide when dealing with events that produce a sudden jump, like instrument replacements or relocations. Regression analysis can be used when looking for artificial trends, like those derived from urbanization, gradual change to irrigated crop fields around station or growing trees producing a shadow.

Several other methods are used to search for breakpoints in data. Usually, this is achieved by performing a set of statistical tests between two data samples composed by consecutive data, moving the contact point between them one element at a time. An *n-sized* dataset can be searched for breakpoints by using a fixed size window (comparing data points *1 to 10* vs. *11 to 20; 2 to 11 vs. 12 to 21; …; n-20 to n-11 vs. n-10 to n*) or by using a varying sized window (comparing data points *1 to 10 vs. 11 to n; 1 to 12 vs. 13 to n;…; 1 to n-11 vs. n-10 to n*). Although repeated statistical tests increase the type-I error probability, each test assesses the likelihood of the last element in the first sample to constitute a breakpoint. Some approaches can detect more than one discontinuity at the same test run some others are designed to find only one at a time. As there certainly can be more than one discontinuity in the data, the firstly labeled point is used then to split the time series in two pieces which are searched again for further discontinuities. For this purpose, the *t-test* or similar formulations based in the evaluation of the change in mean have been widely used. Other approaches adjust a regression line to the data before and after the year being tested and evaluate the change in slope. Finally, some techniques are based in the use of rank order change point detection, like the Wilcoxon-Mann-Whitney test. This particular approach is advisable when the normality of the data is in doubt. Normality is usually more difficult to ensure when dealing with precipitation, and is always more easily achieved in year averaged or accumulated quantities than in monthly data. In such cases it can be helpful to apply a data transformation, like cube roots, before homogenizing to achieve normality.

Some techniques only run the test once, trusting the reference to be homogeneous, while others engage in an iterative procedure in which all the stations in the data set are seen consecutively as candidates and references. This is done to produce some preliminary homogenized data, which will be used in the final homogenization process. When analyzing data from a station, we have to keep in mind that, if we run the selected test on the 12 monthly series, it is possible to find different breakpoints in each one. This is quite understandable, due to the randomness of the time series and because it is obvious that some causes of inhomogeneity can have a larger impact in summer than in winter or the other way around. Same thing may happen when comparing day-time and night-time temperatures. For this reason, it is good to start the analysis over averaged quantities like annual means, which have also less year-to-year variability and usually allow a better detection. In some cases, like temperature, it is recommended to take seasonally averaged data instead of annual means to account for opposed summer-winter effects. It also needs to be mentioned that detection tests have less power in detecting breakpoints near the start and end of a series.

Table 2 lists a series of different approaches to homogenization developed and applied by different groups/authors and Figure 12 shows the different steps to be taken for the homogenization of monthly to annual time series.

Table 2: Review of different widely used techniques for inhomogeneity detection and homogenization. Descriptions and references are obtained from THOMAS C. PETERSON et *al.* (1998), except those marked with *

METHOD	DESCRIPTION
BUISHAND RANGE TEST* Buishand, T.A.. 1982. 'Some methods for testing the homogeneity of rainfall records. *Journal of Hidrology* **58**: 11-27.* Wijngaard, J.B., Klein Tank, A.M.G. and Können, G.P. 2003: 'Homogeneity of 20th century European daily temperature and precipitation series'. *Int. J. Climatol.*, **23**: 679-692.*	The Buishand range test is defined as $S_k^* = \sum_{i=1}^{k}(Y_i - \overline{Y})$ where $k = 1, \ldots, n$. When a series is homogeneous the values of will fluctuate around zero, because no systematic deviations of the Yi values respect to their mean will appear. If a break is present in* year K, then S_k^* reaches a maximum (negative shift) or a minimum (positive shift) near the year $k = K$. The significance of the shift can be tested $R = \left(\max_{0 \le k \le n} S_k^* - \min_{0 \le k \le n} S_k^*\right) / s$, where s is the standard deviation of S_k^*. Buishand (1982) gives ccritical values for the test.
CAUSSINUS-MESTRE TECHIQUE Caussinus, H. and Lyazrhi, F. 1997. 'Choosing a linear model with a random number of change-points and outliers', *Ann. Inst. Stat. Math*., **49**, 761–775. Caussinus, H. and Mestre, O. 1996. 'New	The Caussinus-Mestre method simultaneously accounts for the detection of an unknown number of multiple breaks and generating reference series. It is based on the premise that between two breaks, a time series is homogeneous and these homogeneous sections can be used as reference series. Each single series is compared to others within the

mathematical tools and methodologies for relative homogeneity testing', Proceedings of the Seminar for Homogenization of Surface Climatological Data, Budapest, 6–12 October, pp. 63–82. Mestre, O. and Caussinus, H., 2001. A Correction Model for Homogenisation of Long Instrumental Data Series, in Brunet, M. and López, D., Detecting and Modelling Regional Climate Change, Springer, pp13-19* Caussinus, H., and Mestre, O. (in press). 'Detection and correction of artificial shifts in climate series'. Submited to JRRS, series C. *	same climatic area by making series of differences (temperature, pressure) or ratio (precipitation). These difference or ratios series are tested for discontinuities. When a detected break remains constant throughout the set of comparisons of a candidate station with its neighbours, the break is attributed to the candidate station time series. Recently, the authors have developed a new technique based in the comparison of several perturbed series instead of comparing a series to an artificial reference. A two factor linear model (time x series) is introduced for all series at any time, and a penalised likelihood procedure to select the best model. To facilitate computation, a stepwise approach is adopted.
CRADDOCK TEST Craddock, J.M. 1979. 'Methods of comparing annual rainfall records for climatic purposes', *Weather*, **34**, 332–346. Auer, I., 1992. Experiences with the Completion and Homogenization of Long-term Precipitation Series in Austria, Centr. Europ. research. initiative, Proj. Gr. Meteorology,	Developed by Craddock (1979), this test requires a homogeneous reference series though sometimes long enough homogeneous sub-periods are sufficient (Boehm, 1992). The Craddock test accumulates the normalized differences between the test series and the homogeneous reference series according to the formula: $s_i = s_{i-1} + a_i*(b_m/a_m)-b_i$ where *a* is the homogenous reference series, *b* is the time series to be

Wp. 1, Vienna.	tested and a_m and b_m are the time series means over the whole period.
EXPERT JUDGEMENT METHODS Jones, P.D., Raper, S.C.B. Bradley, R.S. Diaz, H.F. Kelly, P.M. and Wigley, T.M.L. 1986a. ‘Northern Hemisphere Surface Air Temperature Variations: 1851–1984’, *J. Climate Appl. Meteorol.*, **25**, 161–179. Rhoades, D.A., and Salinger, M.J. 1993. ‘Adjustment of temperature and rainfall records for site changes’, *Int. J. Climatol.*, **13**, 899–913.	Judgement by an experienced climatologist has been an important tool in many adjustment methodologies because it can modify the weight given to various inputs based on a myriad of factors too laborious to program. For example, when viewing a graphical display revealing a station time series, a neighbouring station time series and a difference series (candidate-neighbour), a subjective homogeneity assessment can factor in the correlation between the stations, the magnitude of an apparent discontinuity compared to the variance of the station time series, and the quality of the neighbouring station’s data along with other information such as the relevance and reliability of the available station metadata. Expert judgement can be particularly helpful in an initial inspection of the stations’ data and when the reliability of certain inputs (e.g. metadata) varies.

INSTRUMENTS COMPARISONS Forland, E.J., Allerup, P., Dahlstrom, B., Elomaa, E., Jonsson, T., Madsen, H., Per, J., Rissanen, P., Vedin, H. and Vejen, F., 1996: Manual for Operational Correction of Nordic Precipitaion Data, DNMI-Repports 24/96 KLIMA, 66 pp. Nichols, N., Tapp, R., Burrows, K., and Richards, D., 1996. 'Historical thermometer exposures in Australia', *Int. J. Climatol.,* **16**. Quayle, R.G., Easterling, D.R., Karl, T.R. and Huges, P.Y., 1991. 'Effects of recent thermometer changes in the cooperative station network', *Bull. Amer. Met. Soc.*, **72**.	Side by side comparisons are useful to derive the impact of instrument substitutions on data homogeneity. They have been used to asses the difference between shielded and non-shielded rain gauges or Stevenson Screens and other stands. Other approaches are based in statistical comparisons of sets of stations using simultaneously different instruments, for example liquid-glass thermometers and maximum-minimum thermistors. In all cases, the goal is to derive correction factors to substract the impact of the instrument substitution on data
MULTIPLE ANALYSIS OF SERIES FOR HOMOGENISATION (MASH) Szentimrey, T. 1996. 'Statistical procedure for joint homogenization of climatic time series', Proceedings of the Seminar for Homogenization of Surface Climatological Data, Budapest, Hungary, pp. 47–62.	The MASH does not assume the reference series are homogeneous. Possible break points and shifts can be detected and adjusted through mutual comparisons of series within the same climatic area. The candidate series is chosen from the available time series and the remaining series are considered as reference series. The role of the various series

Szentimrey, T. 1999: 'Multiple Analysis of Series for Homogenization (MASH)', Proceedings of the Second Seminar for Homogenization of Surface Climatological Data, Budapest, Hungary; WMO, WCDMP-No. 41, pp. 27-46.* Szentimrey, T. 2000: 'Multiple Analysis of Series for Homogenization (MASH). Seasonal application of MASH (SAM), Automatic using of Meta Data', Proceedings of the Third Seminar for Homogenization of Surface Climatological Data, Budapest, Hungary. *	changes step by step in the course of the procedure. Depending on the climatic elements, additive or multiplicative models are applied. A new multiple break points detection procedure has been developed which takes the problem of significance and efficiency into account. This test obtains not only estimated break points and shift values, but the corresponding confidence intervals as well. A special part of the method is appropriate to homogenize the monthly, seasonal and annual series together. The developed version makes possible to use some metadata information – in particular the probable dates of break points – automatically.
MULTIPLE LINEAR REGRESSION Gullett, D.W., Vincent, L. and Malone, L.H. 1991. Homogeneity Testing of Monthly Temperature Series. Application of Multiple-Phase Regression Models with Mathematical Change points, CCC Report No. 91–10. Atmospheric Environment Service, Downsview, Ontario. 47 pp. Vincent, L. 1998. 'A technique for the identification of inhomogeneities in Canadian tem-	The technique is based on the application of four regression models to determine whether the tested series is homogeneous, has a trend, a single step, or trends before and/or after a step. The dependent variable is the series of the tested station and the independent variables are the series of a number of surrounding stations. To identify the position of a step, the third model is applied successively for different locations in time, and the one providing the minimum residuals sum of squares represents the most probable position

perature series', *J. Climate*, **11**, 1094–1104.	in time of a step in the tested series. The procedure consists of the successive application of the four models (Vincent, 1998).
PETTIT TEST* Pettit, A.N. 1979. 'A non-parametric approach to the change-point detection'. *Applied Statistics* **28**: 126-135.* Wijngaard, J.B., Klein Tank, A.M.G. and Können, G.P. 2003: 'Homogeneity of 20th century European daily temperature and precipitation series'. Int. J. Climatol., 23: 679-692.*	This test is a non-parametric rank test. The ranks $r_1, \ldots, r_n$ of a time series $Y_1, \ldots, Y_n$ are used to calculate the statistics $X_k = 2\sum_{i=1}^{k} r_1 - k(n+1)$ where $k = 1, \ldots, n$. If a break occurs in the year E, the statistic is maximal or minima near the year $k = E$., then $X_E = \max_{1 \le k \le n} \lvert X_k \rvert$. Significance tables are provided by Petit (1979)
POTTER'S METHOD Plummer, N., Lin, Z. and Torok, S. 1995. 'Trends in the diurnal temperature range over Australia since 1951', *Atmos. Res.*, **37**, 79–86. Potter, K.W. 1981. 'Illustration of a new test for detecting a shift in mean in precipitation series', *Mon. Wea. Rev.*, **109**, 2040–2045.	Is a likelihood ratio test between the null hypothesis that the entire series has the same bivariate normal distribution and the alternate hypothesis that the population before the year being tested has a different distribution than the population after the year in question. This bivariate test closely resembles a double mass curve analysis. One part of the test statistic depends on all points on a time series while another part depends only on the points proceeding the year in question. The highest value of the test statistic will be in the year preceding a change in the mean of the candi-

	date station time series. Potter (1981) applied this technique to ratio series of candidate station's precipitation and a composite reference series
RADIOSONDE DATA* Free, M., Durre, I., Aguilar, E., Seidel, D. Peterson, T.C., Eskridge, R.E., Luers, J.K. Parker, D., Gordon, M., Lanzante, J., Klein, S., Christy, J., Schroeder, S., Soden, B., McMillin, L., and Weatherhead, E., 2001: 'Creating Climate Reference Datasets. CARDS Workshop on Adjusting Radiosonde Temperature Data for Climate Monitoring' , *Bull. Amer. Meteor. Soc.* **83**, 891-899. *	During the past 60 years, radiosondes have been launched around the globe to collect information on vertical profiles of temperature, humidity and other atmospheric variables. The radiosonde record provides more detailed vertical resolution and a longer history than the satellite record. Archived time series of radiosonde measurements can often be plagued by inhomogeneities that compromise the validity of trends calculated from data. Currently, various groups are working to identify and remove these inhomogeneities to make the data more suitable for climate studies. The adjustment of radiosonde data also requires the identification of artificial discontinuities in the data, estimation of the size of these discontinuities, and application of adjustments. But due the special characteristics of upper air data, special strategies have to be adopted. The reference quoted in this table provides insights on different approaches and further refer-

	ences.
RANK-ORDER CHANGE POINT TEST Siegel, S. and Castellan, N. 1988. Nonparametric Statistics for the Behavioural Sciences, McGraw-Hill, New York, 399 pp. Lanzante, J.R. 1996. 'Resistant, robust and nonparametric techniques for the analysis of climate data. Theory and examples, including applications to historical radiosonde station data', *Int. J Climatol.*, **16**, 1197–1226.	Using a test based on the ranks of values from a time series has the benefit that it is not particularly adversely affected by outliers. Lanzante (1996) describes such a non-parametric test related to the Wilcoxon-Mann-Whitney test. The test statistic used is computed at each point based on the sum of the ranks of the values from the beginning to the point in question (Siegel and Castellan,1988). And the maximum value is considered the point of a possible discontinuity.
STANDARD NORMAL HOMOGENEITY TEST Alexandersson, H. and Moberg, A. 1997. 'Homogenization of Swedish temperature data. Part I: A homogeneity test for linear trends', *Int. J. Climatol.*, **17**: 25–34. Alexandersson, H. 1986. 'A homogeneity test applied to precipitation data', *J. Climate*, **6**, 661–675.	Is a likelihood ratio test. The test is performed on a ratio or difference series between the candidate station and a reference series. First this series is normalized by subtracting the mean and dividing by the S.D. In its simplest form, the SNHT's statistic is the maximum of $Tv = v(\overline{z}_1)^2 + (n - v)(\overline{z}_2)^2$, where $\overline{z}_1$ is the mean for the series from data point 1 to v and $\overline{z}_2$ is them mean of the series from *v+1* to the end, *n*. There are now variations in this test to account for more than one

Hanssen-Bauer, I., Forland, E. 1994: ‘Homogenizing long Norwegian precipitation series’, *J. Climate*, **7**, 1001-1013.	discontinuity, testing for inhomogeneous trends rather than just breaks, and inclusion of change invariance
STOP-TREND METHOD* Kobysheva, N and Naumova, L. 1979: Works of the Main Geophysical Observatory, 425, Saint Petersburg, Russia.*	A non-parametric test. Data are sorted by date and consecutive ranks are assigned. Then, the time series is split into $k = n^{0.5}$, where n is the number of observations, with size $l=(max-min)/k$., where *max* and *min* are the maximum and minimum value in the dataset. Inside each interval, if the difference between consecutive ranks exceeds a critical level based on the Kolmogorov coordination criterion, the observation corresponding to the first rank is labelled with and “A” and the second with a “B”. Once all the intervals have been evaluated, if two adjacent observations in the overall time series are flagged with “A” and “B”, the “A” observation defines a breakpoint.
TWO-PHASE REGRESSION	Solow (1987) described a technique for detecting a change in the trend of a time series by identifying the change point in a two

Solow, A. 1987. 'Testing for climatic change: an application of the two phase regression model', J. Climate Appl. Meteorol., 26, 1401–1405. Easterling, D.R. and Peterson, T.C. 1995a. 'A new method for detecting and adjusting for undocumented discontinuities in climatological time series', Int. J. Climatol., 15, 369–377. Easterling, D.R. and Peterson, T.C. 1995b. 'The effect of artificial discontinuities on recent trends in minimum and maximum temperatures', *Atmos. Res.*, **37**, 19–26.	phase regression where the regression lines before and after the year being tested were constrained to meet at that point. Since changes in instruments can cause step changes, Easterling and Peterson (1995a,b) developed a variation on the two-phase regression in which the regression lines were not constrained to meet and where a linear regression is fitted to the part of the (candidate-reference) difference series before the year being tested and another after the year being tested. This test is repeated for all years of the time series (with a minimum of 5 years in each section), and the year with the lowest residual sum of the squares is considered the year of a potential discontinuity.

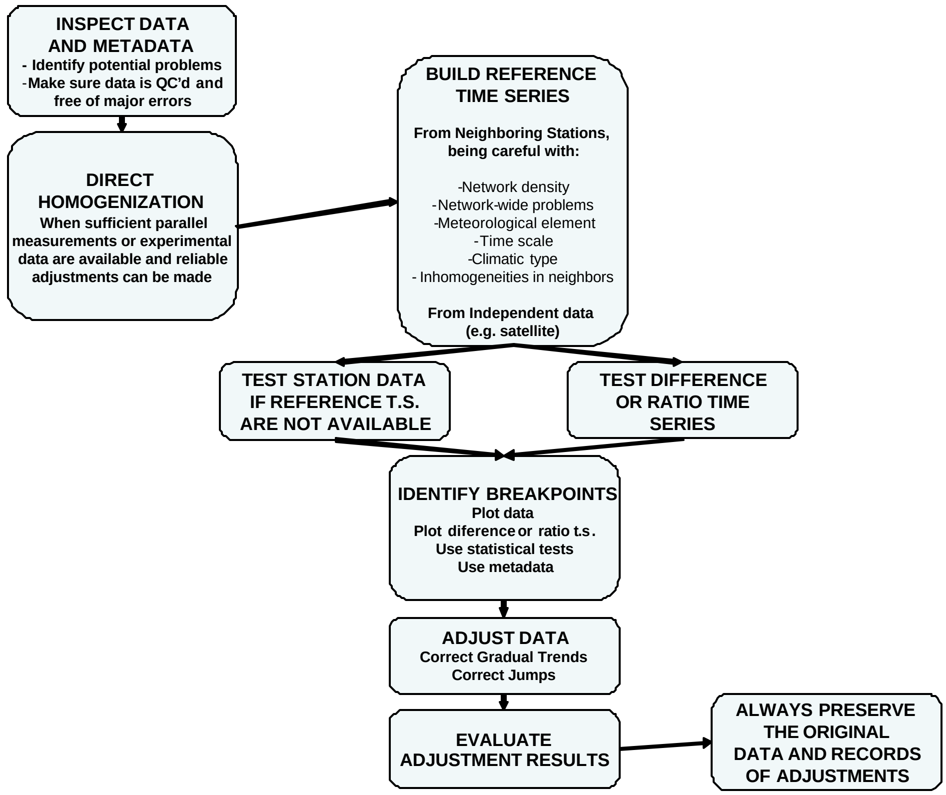


Figure 12: Schematic representation of an approach to homogenization procedures for monthly to annual climate records

Principal component analysis: Section (3.2.4) in this chapter gives details about principal component analysis, the primary objective of which is to reduce the dimensionality in a dataset. Even though it is not a traditional method for homogeneity testing, its ability to draw functions from a dataset that together describe the majority of variance information without repetition could be used to estimate the homogeneity of a dataset. In applying principal component analysis in this sense the procedure is able to separate the signal (which contains useful information about the climate entity) and the noise (which contains no information about the climate entity measured). Figures 13 presents time series plots of annual averaged temperature and precipitation and the mean annual cycles of temperature and precipitation for the NCEP reanalysis data from NCEP. The error is the difference between the original dataset and the filtered dataset. The differences are very small because in this case, it is the difference between the methods of homogeneity testing utilized by the supplying institution and principal component analysis and not the error in the dataset itself as the reanalysis data has already undergone homogeneity testing by their supplying institutions.

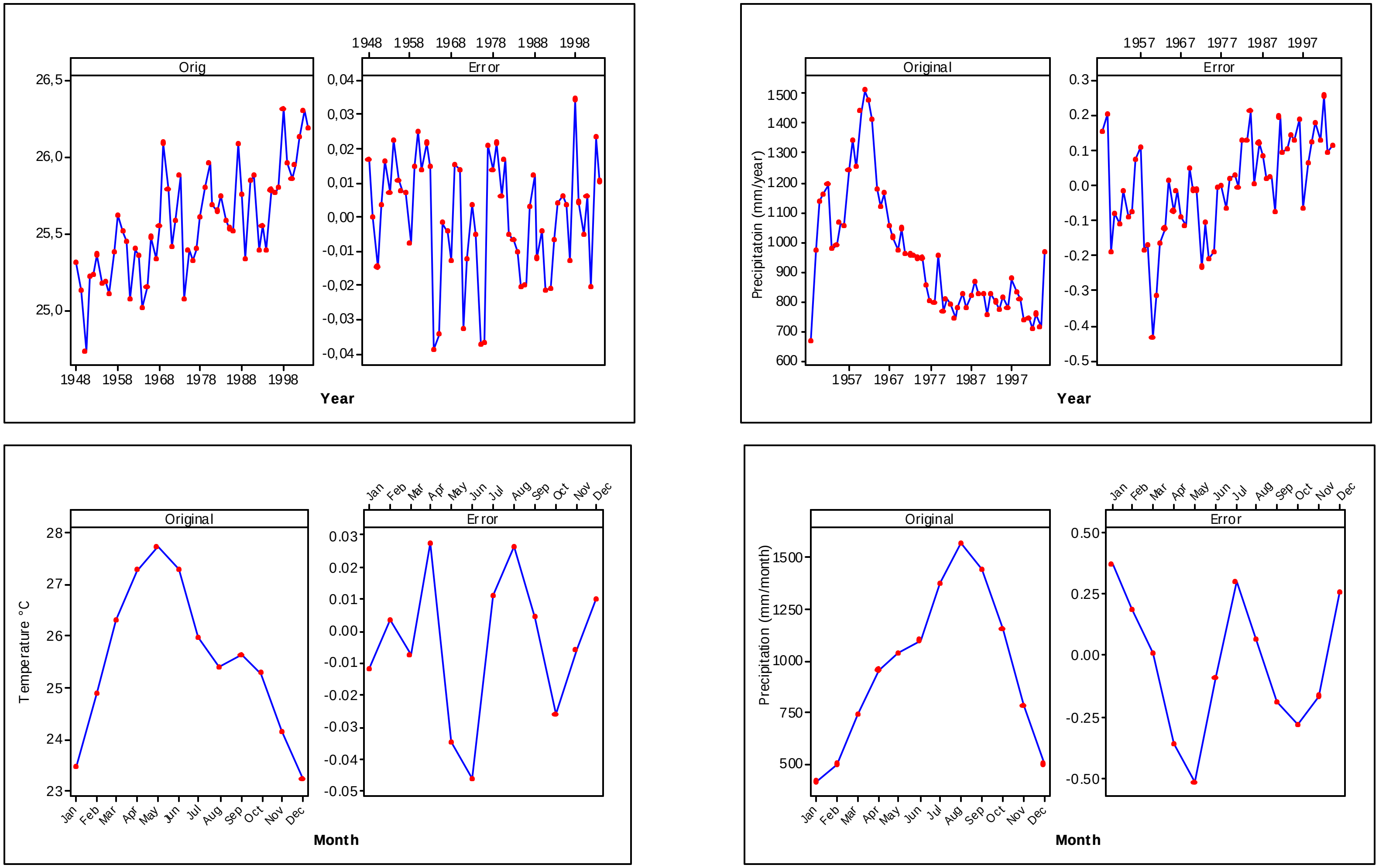


Figure 13: Time series plot of annual mean temperatures/ precipitation (top), annual cycle of temperature / precipitation (bottom) 1948-22004

3.3 Analyses methodology

The goal of this study is to identify trends in temperature and precipitation during the 2nd half of the 20th century and to identify patterns of precipitation over Central Africa. The linear regression method and principal component analysis seem most compatible with these objectives. All graphs in this study were created with Minitab software and maps displayed in this study were generated using the kriging interpolation method with ArcGIS spatial analyst. The kriging method has an advantage over other interpolation techniques because it incorporates the spatial correlation of data while other interpolation techniques do not. The applied methods for analysis are detailed below.

3.3.1 Linear regression

An experimental study of the relation between two variables is often motivated by a need to predict one from the other. Regression is a statistical technique that allows decision makers not only to establish quantitative relationships among such variables but also measure the “strength” of the relationship. In regression analysis a mathematical equation is developed which relates an unknown variable to a known quantity of interest. The known variable(s) is (are) called the independent (explanatory, predictor) variable(s) x, while the variable to be predicted is the dependent (or response) variable y. The object is to find the nature of the relation between x and y from experimental data and use the relation to predict the response variable y from the input x. Naturally, the first step in such a study is to plot and examine the scatter diagram. If a linear relation emerges, the calculation of the numerical value of r will confirm the strength of the linear relation. Its value indicates how effectively y can be predicted from x by fitting a straight line to the data. The coefficient of determination, or r^2, expresses the strength of the relationship between the X and Y variables. It is the proportion of the variation in the Y variable that is "explained" by the variation in the X variable. r^2 can vary from 0 to 1; values near 1 mean the Y values fall almost right on the regression line, while values near 0 mean there is very little relationship between X and Y

A line is determined by two constants: its height above the origin (intercept) and the amount that y increases whenever x is increased by the unit (slope). See Figure 14. This best fit line or least square line is close to the points graphed in the scatter plot in terms of minimizing the amount of vertical distance.

$$\hat{y} = \hat{b}_0 + \hat{b}_1 x \tag{1}$$

$$\text{Slope } \hat{b}_1 = \frac{S_{xy}}{S_{xx}} = \frac{\sum (x-\overline{x})(y-\overline{y})}{\sum (x-\overline{x})^2} \tag{2}$$

$$\text{Intercept } \hat{b}_0 = \hat{y} - \hat{b}_0 \overline{x} \tag{3}$$

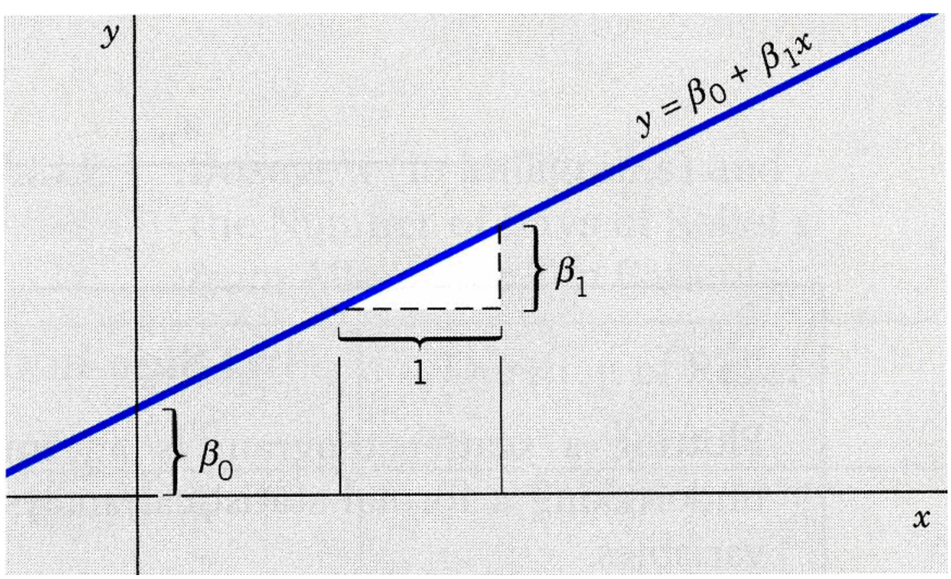


Figure 14: The line $\hat{y} = \hat{b}_0 + \hat{b}_1 x$

Trends in time-series data were analyzed using simple linear regression. The slope bx indicates the average rate of change in the climatic parameter over the time period. One advantage of this method is that it is easy to apply to a large number of sites. A disadvantage is that it can fail to detect trends that are nonlinear but still monotonic (generally in one direction). Other methods, such as the Mann-Kendall test, could be used to detect trends that are monotonic but not necessarily linear, but these only indicate the direction, and not the magnitude trends.

3.3.2 The Kendall's Tau method

To appropriately estimated changes over time, a method of assessing trend behaviour is required. Kendall's Tau (Sen, 1968; Kotz and Johnson, 1983; Kunkel et al, 1999a; Lins and Slack, 1999) is a test for both linear and non-linear trends that makes no strong assumption (i.e. it is nonparametric) about the data concerned. These are useful properties when dealing with potentially non-linear extremes of climate i.e. it relies on the connection of all data pairs (e.g. y_1 and y_2, y_2 and y_3,....)Kendall's Tau is given by the following equation

$$t = \frac{numberofpositiveslopes - numberofnegativeslopes}{total\ number\ of\ segments} \tag{4}$$

If the majority have a downward (upward) gradient, the data is considered to possess a negative (positive) trend. If Kendall's Tau reaches zero, there is no trend evident in the data. Values nearing 1 and -1 imply a vertical distribution (x is constant, y varies) to the data

3.3.3 Significance and stability testing

All trends considered in this study are tested for 'significance'. In statistics, a result is called significant if it is unlikely to have occurred by chance. In traditional frequentist statistical hypothesis testing, the significance level of a test is the maximum probability, assuming the null hypothesis, that the statistic would be observed. Hence, the significance level is the probability that the null hypothesis will be rejected in error when it is true (a decision known as a Type I error, or "false positive"). The significance of a result is also called its p-value; the smaller the p-value, the more significant the result is said to be. Significance is usually represented by the Greek symbol, a (alpha). Popular levels of significance are 5%, 1% and 0.1%. If a test of significance gives a p-value lower than the a-level, the null hypothesis is rejected. Such results are informally referred to as 'statistically significant'.

3.3.4 Principal component analysis

One goal of statistical analysis of climate data is to extract physically meaningful patterns of variability from measurements. The most common methods used to study patterns of climatic parameters over an area are Lund’s (1963) spatial correlation technique, which uses linear correlation between patterns to assess their similarity as well as principal component analysis a variant of factor analysis which has been used e.g. by Kutzbach (1970) and Kidson (1975) to analyze global patterns of variability. Lund’s method was specifically developed as an objective technique for describing and classifying patterns and has been commonly used as a pattern recognition device for identifying characteristic configuration of meteorological variables (Blasing and Lorgren, 1980) and measuring the persistence of meteorological fields (Horel, 1985). In contrast the principal component technique was created primarily to reduce dimensionality in a dataset, (Richman, 1981) although it has been routinely applied to the problem of delineating spatial fields.

Wallis (1965) and Richman (1981) describe the various methods as well as the advantages and limitations of their applications in detail. One disadvantage of the correlation method is that the derived spatial patterns are not orthogonal and hence of little value in deriving predictive equations (Blasing, 1975). One draw back in the principal component technique is that the associated spatial patterns are mathematical devices which except for the first principal component are dictated by the constraint of orthogonality. Consequently, the principal components do not necessarily lend themselves to physical interpretation and may bear little resemblance to synoptic meteorological fields.

The application of any of the methods is more problematic for analyzing rainfall than temperature or pressure because rainfall exhibit a high degree of spatial and temporal variability, is significantly influenced by small scale factors and is prone to large observational errors. In view of the complexity of Lund's correlation method particularly with regards to its requirement of several arbitrary decisions, the current research will apply the principal component technique to study patterns of rainfall variability over Central Africa.

Principal component analysis is a methodology in wide usage by the climate community that allows for the spatial and temporal analysis of the variability of physical fields (Preisendorfer, 1988; Wilks, 1995; Esteban-Parra et al., 1998; Rodriguez-Puebla et al., 1998). Researchers have utilized PCA (also termed Empirical Orthogonal Functions, or EOF) to investigate centers of variance in a number of data (Maheras et al., 1999a; Serrano et al., 1999; Quadrelli et al., 2001; Bordi and Sutera, 2002), but its main function is to reduce dimensionality of a data set (Preisendorfer, 1988; Jolliffe, 1990). The PCA method can be used to study the climatology of the Central African region (through centers of variance) in terms of precipitation patterns and in an attempt to include otherwise unwieldy multi-dimensional data as model predictors representing those regimes .

If a data set contains points that are substantially correlated then it also contains redundant information. Atmospheric fields tend to display a large amount of redundancy as concurrent points are often interdependent and they are therefore well suited to this form of treatment PCA seeks to remove this redundancy by drawing functions from a dataset that together describe the majority of variance information without repetition. These functions are the 'principal components' of the data. This leaves the researcher with a greatly reduced data set that still accounts for the majority of variance.

The principal components are given by the eigenvectors of the covariance matrix of the data set anomalies i.e., the data set with the mean subtracted from the original. The variance-covariance matrix is the anomaly data (Eqn. 6), multiplied by its own transpose and divided by *n-1* (Eqn 7), where *n* is the number of observations, or row-length in the case of 2-D matrices.

$$x' = x - \overline{x} = \begin{pmatrix} x_1 \\ x_2 \\ . \\ . \\ . \\ x_n \end{pmatrix} - \begin{pmatrix} \overline{x_1} \\ \overline{x_2} \\ . \\ . \\ . \\ \overline{x_n} \end{pmatrix} \tag{6}$$

$$[S] = \frac{x'x'^T}{n-1} \tag{7}$$

The result is a square matrix $[S]$ whose diagonal represents the sample variances and whose other elements express inter-row covariance. Eigenvectors *(e)* are those non–trivial vectors that can be multiplied by an eigenvalue l to give the same result (Eqn. 8).

$$[S]e = l\,e \tag{8}$$

In most cases there will be as many eigenvectors as there are rows and columns in the original matrix, and in the case of symmetric matrices (such as the covariance matrix) each eigenvector of the matrix is orthogonal to every other eigenvector. This is the property that ensures successive principal components explain exclusive proportions of the variance. The equation for each (mth) principal component U_m is then:

$$U_m = e_m^T x' = \sum_{i=1}^{n} e_{nm} x'_n \qquad n = 1, \ldots, N \tag{9}$$

After a geophysical field that varies with time (a three dimensional matrix) is subjected to PCA, two useful products are available i.e. a number of principal component score time series (PCs) equal in quantity to the number of matrix elements in the field (without time) that are orthogonal to each other and a set of weightings (or 'loadings') for each PC that numerically describes the way a given mode of variance affects each field element, and can then be plotted in map format to display centres of variance. These loadings are simply the elements of each eigenvector resulting from the variance-covariance matrix detailed above.

The loading plots can be interpreted directly as modes of variance in the field, but care must be taken to remember that each mode is orthogonal to every other mode (as they are eigenvectors). Large loading anomalies can be taken to represent areas where the relevant PC is strong, and thus where orthogonal modes of variance are the most apparent in the field's behaviour. Where these structures resemble a known phenomenon variance can sometimes be attributed to those processes. If a persistent loading anomaly for a pressure PC is apparent over the Azores, for example, then that indicates high levels of relative variability in that region. If the shape of the anomaly is consistent with the southward centre of the North Atlantic Oscillation then the PC under consideration can be assumed to represent some measure of the NAO phenomenon. If a loading anomaly pattern displays no apparent relationship to known phenomena, and displays unlikely characteristics (such as opposing poles of variance in each of the corners of the region), it may be rejected as an artefact of orthogonality (Priesendorfer 1988; Wilks, 1995). Such artefacts are generally more common for low variance components than for high variance components (Wilks, 1995).

The fall-off of variance explained with each successive PC tends to be roughly exponential. PCs can be rejected when they fall below a certain proportion of variance or lie below a level given by the 'Scree test'. The Scree test simply plots the amount of variance explained for each component in succession and gives the level where the curve starts to flatten as a cut-off point. Figure 16 gives an example of a scree plot from PCA analysis of a precipitation dataset. From the illustrated figure, four principal components will be selected if the selection criterion is based on the scree plot test. The selected PCs combined explain 65.3 % of the total variance in the precipitation dataset, Table 3. In this study components are rejected if they fall below the Scree test cut-off point or if the amount of variance explained is less than 0.04 (Briffa et al., 1983). The use of both methods ensures a high total explained variance (the variance explained by all retained components taken together), while ensuring that the number of retained components is low. It is the process of rejecting components that reduces the dimensionality of a given atmospheric field, and allows the large amounts of data associated with atmospheric fields (that vary with time) to be included as model predictors.

Table 3: Results of PCA analysis of precipitation dataset showing the eigenvalues and percentage variance and cumulative variance of the resultant principal components

	PC1	PC2	PC3	PC4	PC5	PC6	PC7	PC8	PC9
Eigenvalue	84.2	27.1	23.9	11.0	9.2	8.3	7.1	5.5	4.8
% variance	37.6	12.1	10.7	4.9	4.1	3.7	3.1	2.4	2.1

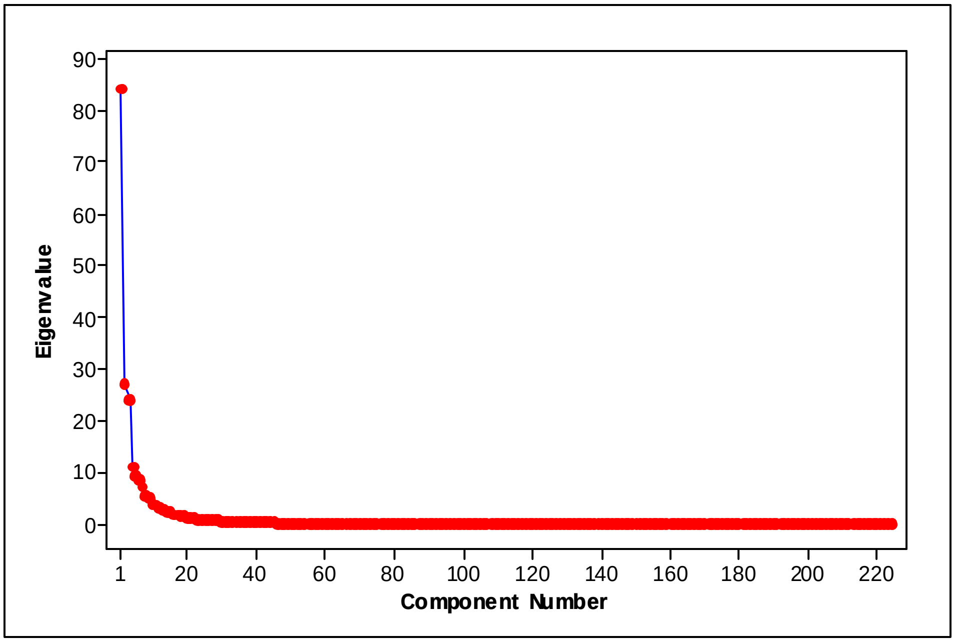


Figure 15: Scree plot from PCA analysis of precipitation data

4 Results and Discussion

4.1 Temperature trends over Central Africa

Annual and decadal average temperature data were derived using daily mean temperatures from the NCEP reanalysis data. The mean temperature for the 57-year period from 1948 through 2004 was 25.6 degrees Celsius (ºC). The warmest decade was the 1998 - 2004, with a mean of 26.1 degrees ºC. The coolest decade was the 1948 -1957, with a mean temperature of only 25.1 ºC.

Table 4: Table of Decadal Temperature Means from 1948 through 2004, and the 57-year mean

Temperature ºC/decade						
1948-1957	1958-1967	1968-1977	1978-1987	1988-1997	1998-2004	57 year mean
25.1	25.3	25.5	25.7	25.6	26.1	25.6

The data of Table 4 indicate there was a significant warming period from the 1948-1957 into the 1978-1987, followed by cooling in the 1988-1997. The 1948 - 1957, 1958 - 1967, 1978 - 1987 and 1988 - 1997 oscillated around the 57 year mean, followed by another warming trend in 1998 - 2004. Temperatures in the 1978 - 1988 were less than a half degree above the 57 year mean, yet significantly cooler than the mean in 1998 - 2004.

Figure 16 presents annual temperature anomalies, i.e. the differences from the mean for the period 1948-2004. The strongest negative temperature anomaly was experienced in 1950 and the strongest positive anomaly happened in 1999, respectively. Most years between 1948 -1968 are characterized by negative anomalies and the last three decades of the 20th century present a dry temperature regime than the previous decades. The estimated temperature change over the 57 year period for Central Africa is 0.15 K / decade. A further analysis of data from area averages over 2.5 degree in latitude and longitude revealed two distinct temperature change zones, see Figure 17. These zones are identified as the Sahel and East Africa and can be found in the Northern and Eastern parts of the study area. The temperature change within these zones is –0.06 to -0.04 K / year and 0.04 to 0.06 K/ year, respectively. Timeseries plots of some grid area averages in the Sahel and East Africa are provided in Figures 18 and 19 respectively. These plots show the relative response of local climate to regional; continental temperature and precipitation change.

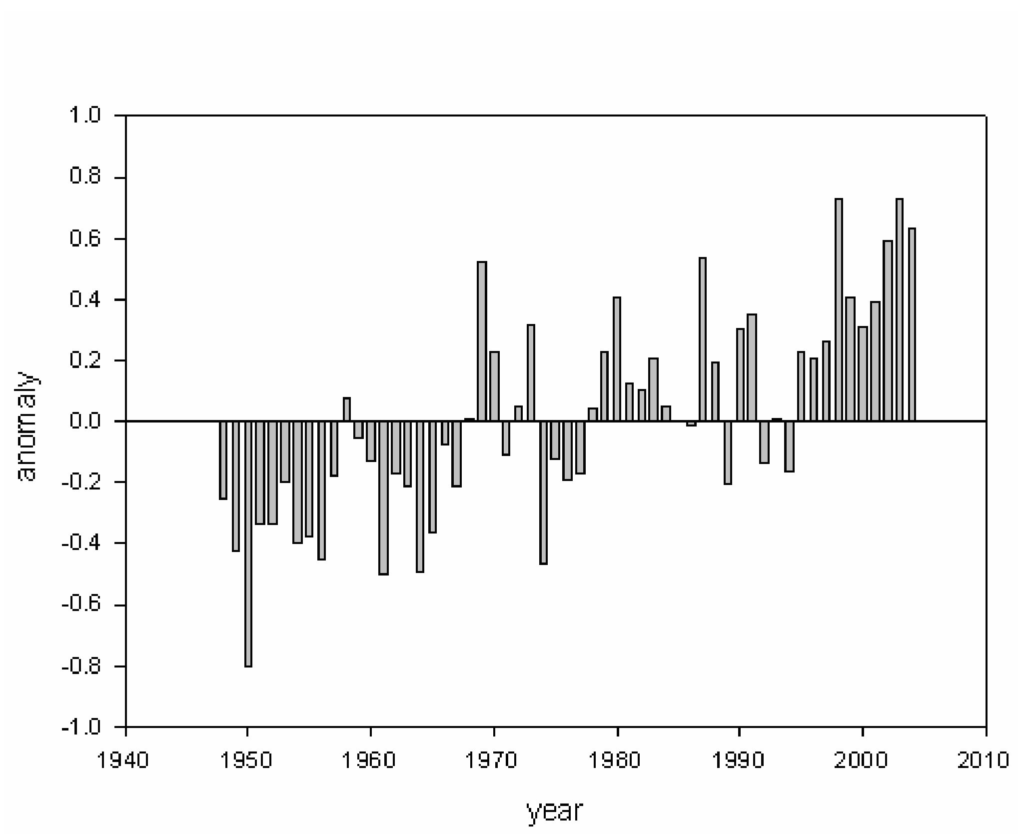


Figure 16: Annual temperature anomalies, the anomalies are calculated on the mean for 1948-2004 from the NCEP reanalysis dataset

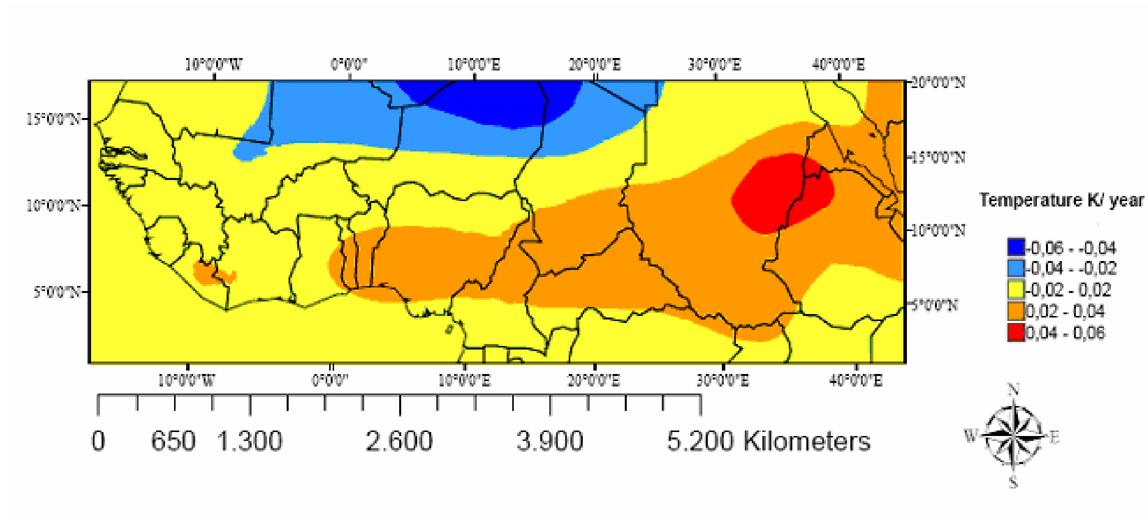


Figure 17: Annual mean temperature trends, K/ year, (1948 to 2004)

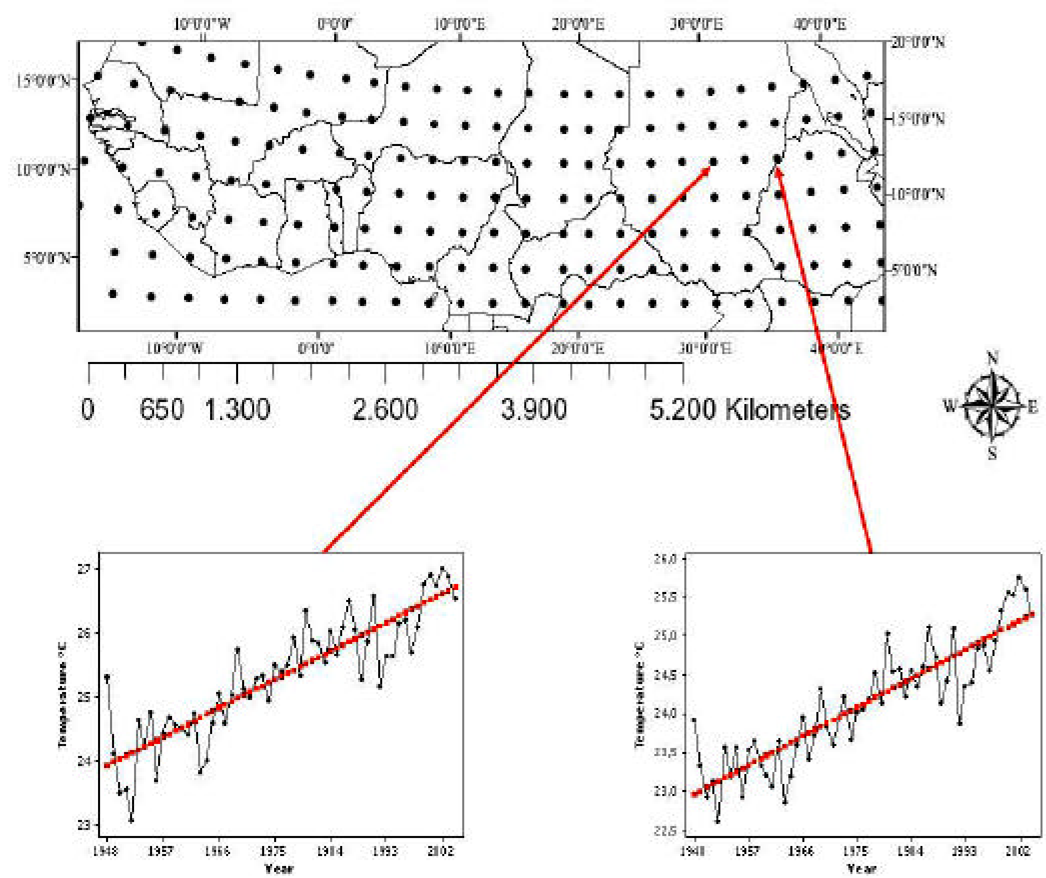


Figure 18: Temperature trends of two grid point averages in East Africa (13 ° N 43.25 ° E and 13 ° N 38.25 ° E)

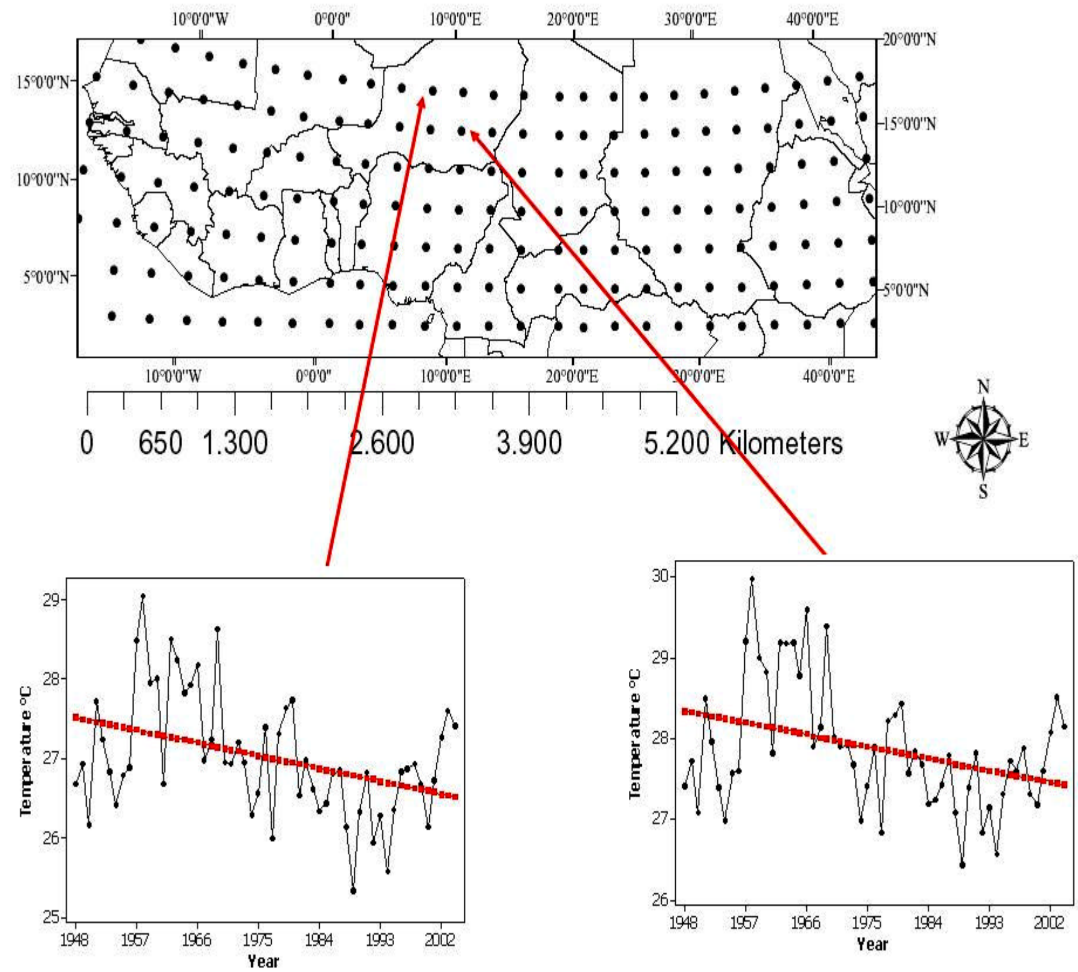


Figure 19: Temperature trends of two grid point averages in Sahel (18 º N 10.75 º E and 15 º N 5.75 º E

A plot of data from three climatic periods 1951-1980 Figure 20, 1961-1990 Figure 21, and 1971-2000 Figure 22 categorised according to the World Meteorological Organisation (WMO) definition indicates that there was no significant temperature change between the first two climatic periods. Both periods experienced a temperature change of 0.16 K / decade with average mean temperatures of 25°C and 26°C respectively. The last period experienced significant cooling with estimated temperature change of 0.13 K / decade.

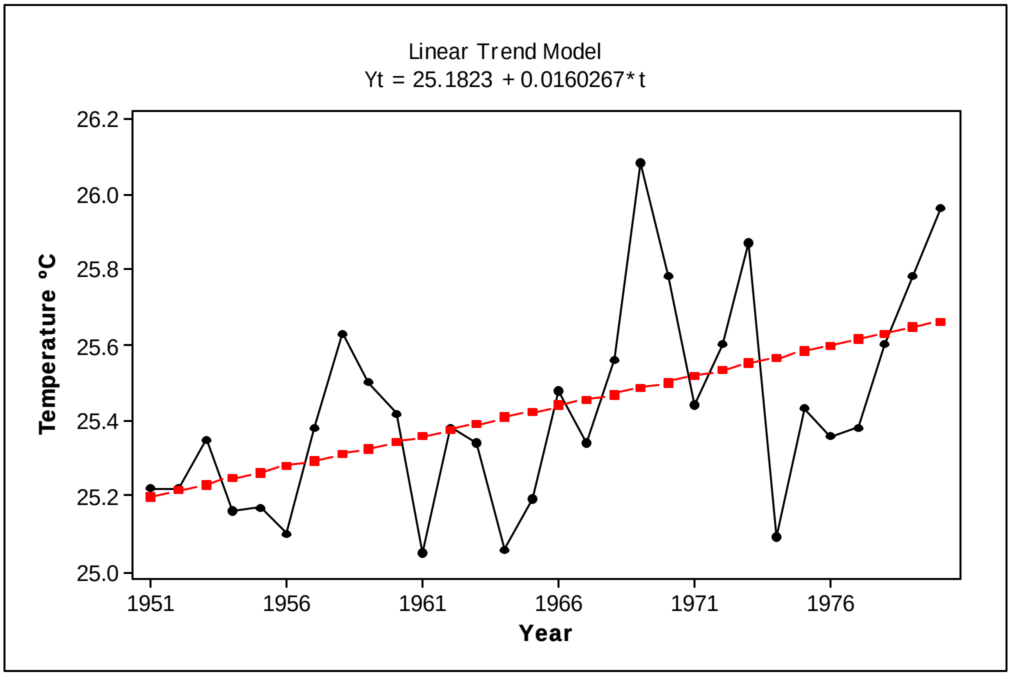


Figure 20: Annual mean temperature trend for first climatic period 1951-1980.

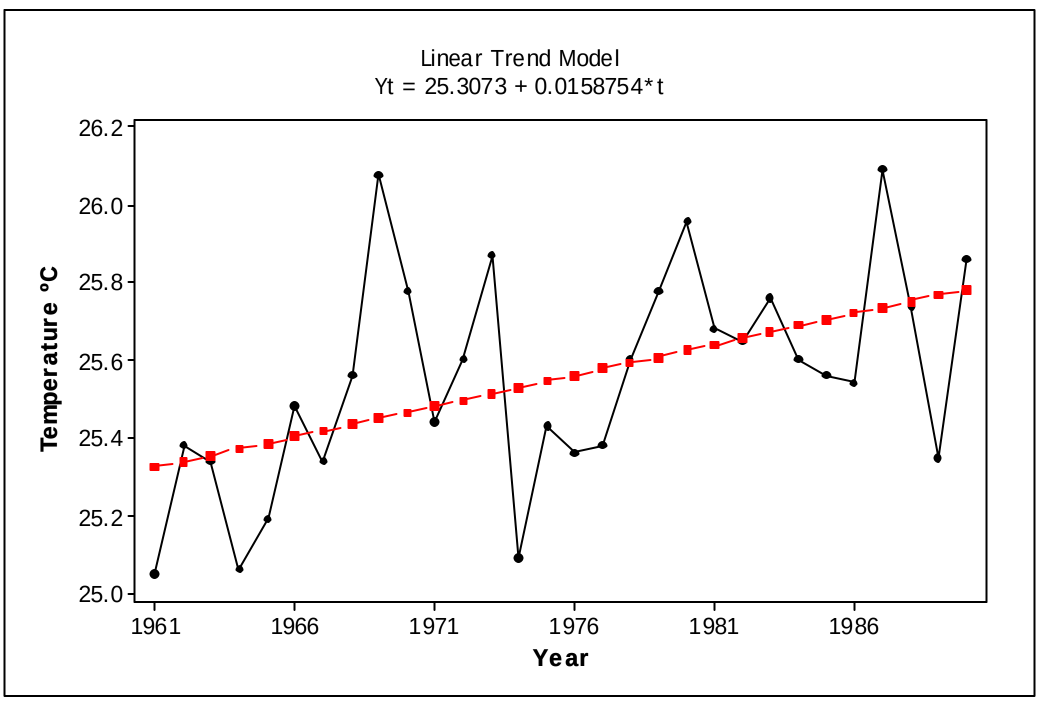


Figure 21: Annual mean temperature trend for second climatic period 1961-1990

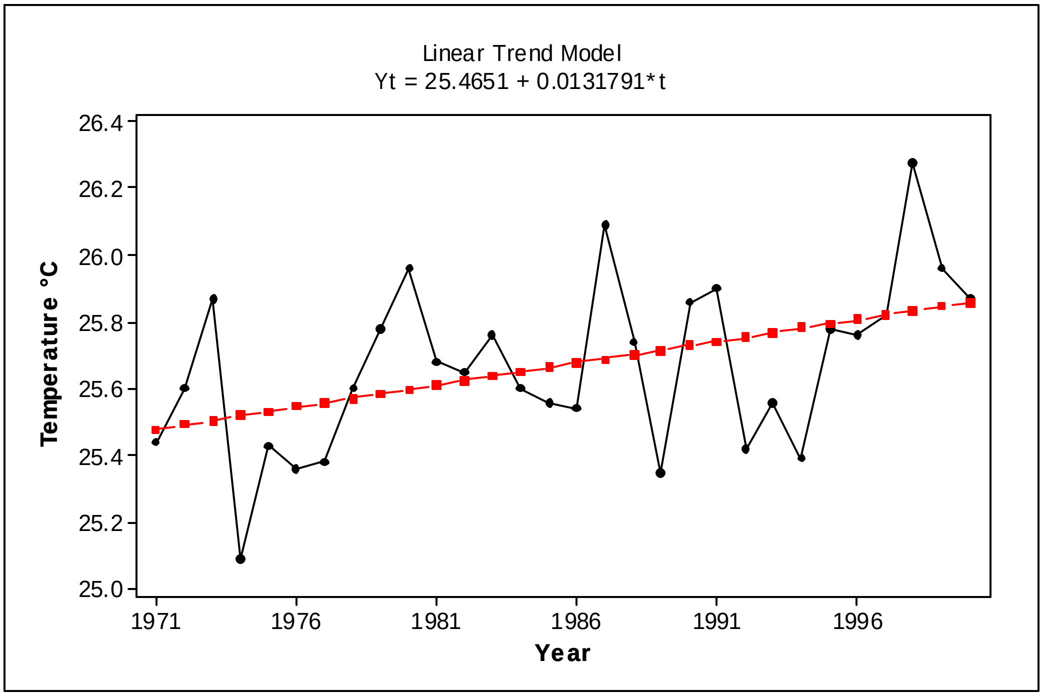


Figure 22: Annual mean temperature trend for third climatic period 1971-2000

The seasonal cycle of rainfall is not synchronized throughout Central Africa due to the movement of the ITCZ across the equator. Rainfall in the Sahel occurs between June and September and in East Africa the main rainfall season occurs between March and May. Figures 23, 24 present temperature trends for raining seasons in the Sahel and East Africa. The seasonal temperatures for the Sahel decreased at an estimated rate of -0.02 K/year. In East Africa temperatures increased by 0.04 K/ year.

The data of Table 5 indicate for the Sahel, 1958-1957 recorded the highest average temperature / decade with a mean of 29.9 ºC and estimates for the last two decades 1988-97 and 1998-2004 were significantly lower than the average mean for the entire Sahel, 29.4 ºC. In East Africa the lowest average temperature/ decade was in 1948-57, average decadal temperatures increased sturdily till 1978-87, then reduced in 1988- 97 and increased till it reached a maximum in 1998 -2004.

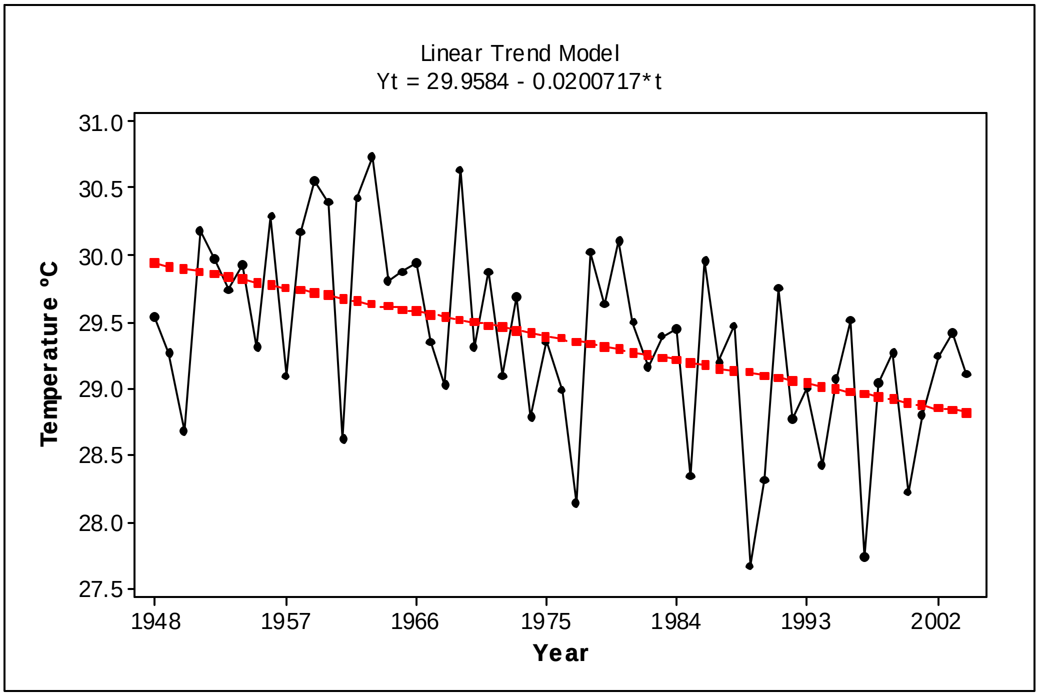


Figure 23: Annual mean June – September raining season temperature trend Sahel

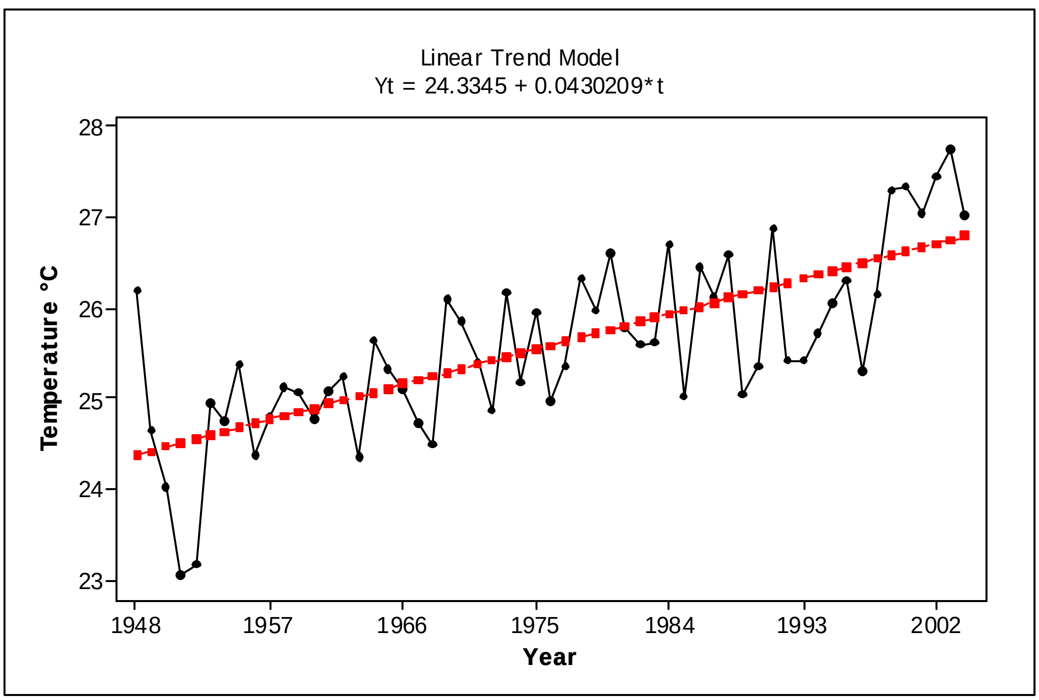


Figure 24: Annual mean March – May temperature trend East Africa

Table 5: Table of decadal temperature means (June to September; March – May) raining seasons from 1948 through 2004, and the 57-year mean for Sahel and East Africa

Temperature ºC/decade							
	1948-57	1958-67	1968-77	1978-87	1988-97	1998-2004	57 year mean
Sahel	29.6	29.9	29.3	29.5	28.8	29.0	29.4
East Africa	24.5	25.0	24.4	26.0	25.8	27.3	25.7

4.2 Precipitation trends over Central Africa

Figure 25 presents annual precipitation anomalies, i.e. the differences from the mean for the period 1948-2004. The strongest negative precipitation anomaly was experience in 1948 and the strongest positive anomaly happened in 1961 respectively. Two clear positive and negative anomaly periods are shown in Figure 25; the years between 1949 - 1970 are characterized by positive and the years between 1971 – 2004 characterized by negative anomalies. The estimated precipitation change over the 57 year period is -91 mm / decade.

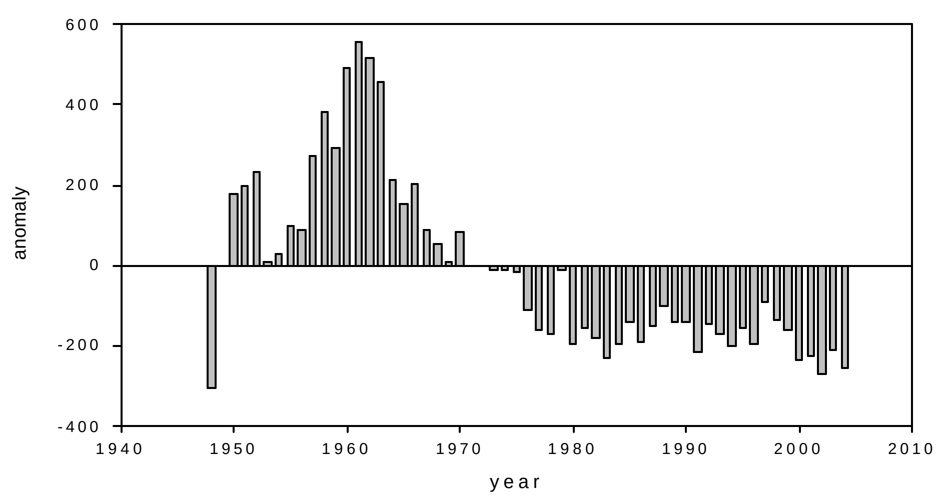


Figure 25: Annual precipitation anomalies, the anomalies are calculated on the mean for 1948-2004 from the NCEP reanalysis dataset.

The data in Table 6 shows that the decadal annual means for Central Africa varied from a low of 664 mm / decade in 1948 – 1957 to a high of 1343 mm / decade in 1958-1967. For the 57 years of data from 1948 through 2004, the mean annual precipitation was 904 mm / decade.

Table 6: Table of Decadal Precipitation Means from 1948 through 2004, and the 57-year mean.

Precipitation mm/ decade						
1948-1957	1958-1967	1968-1977	1978-1987	1988-1997	1998-2004	57 year mean
664	1343	1018	794	864	744	904

An interesting trend appears in the decadal precipitation changes from 1958-1967 through 1998-2004. A trend line would indicate a nearly steady increase in precipitation from one decade to the next. There was an increase in rainfall during 1988-1997; however, given the wide variation in precipitation from one decade to another, this apparent trend may only be coincidental.

From the analysis of data from grid area averages revealed two distinct precipitation change zones, the Sahel and East Africa in the Northern and Eastern parts of the study area, Figure 26 . The estimated precipitation change in the Sahel and East Africa is 6.3 to 12.6 mm / year and –50.8 to – 50.5 mm / year and respectively. Timeseries plots of some grid point area averages of precipitation in the Sahel and East Africa are provided in Figures 27 and 28.

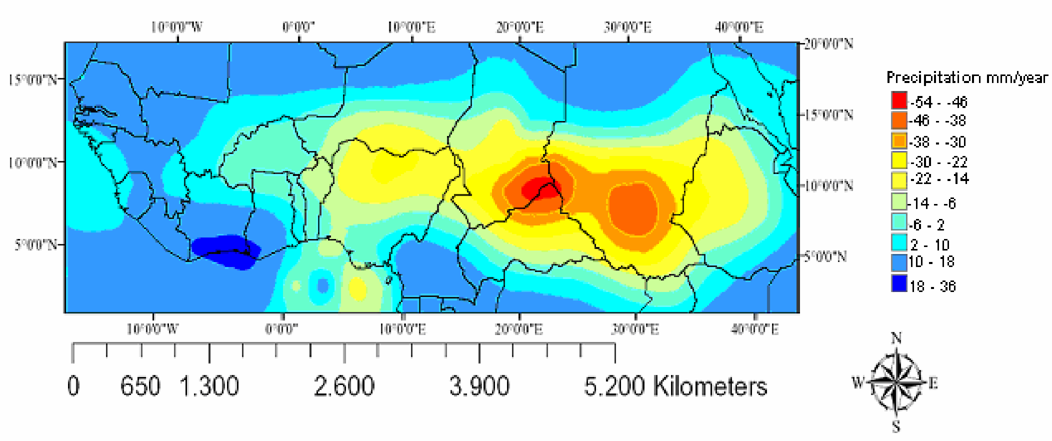


Figure 26: Calculated annual mean precipitation trends (1948 to 2004)

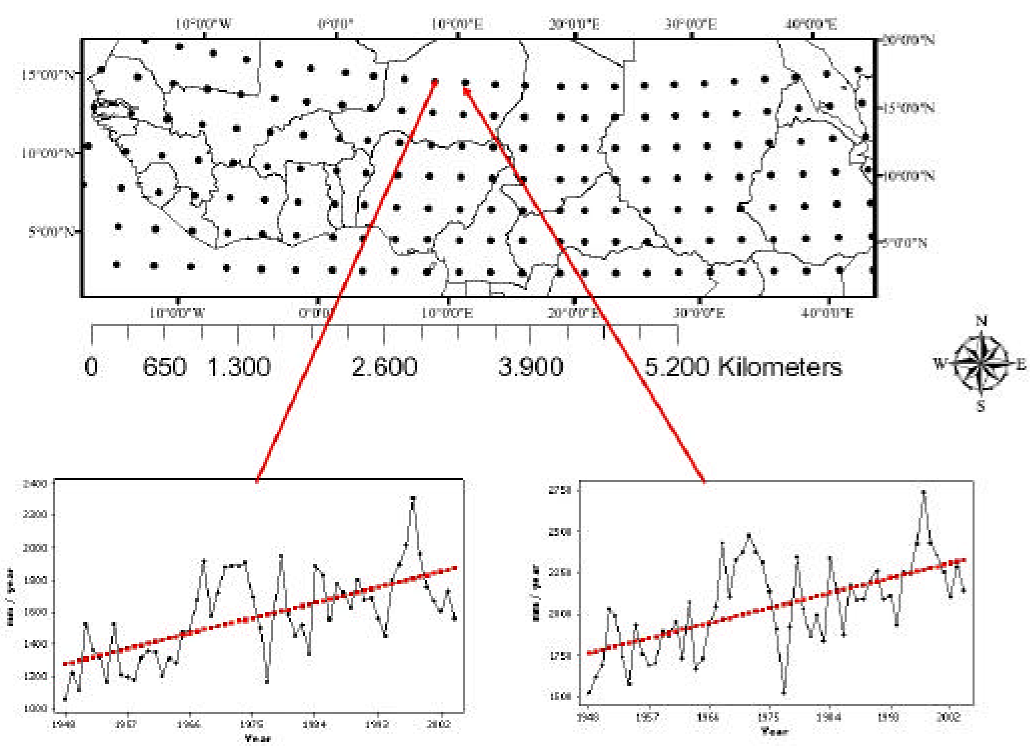


Figure 27: Sahel trends of two grid point averages in Sahel (18 º N 10.75 º E and 18 º N 5.75 º E)

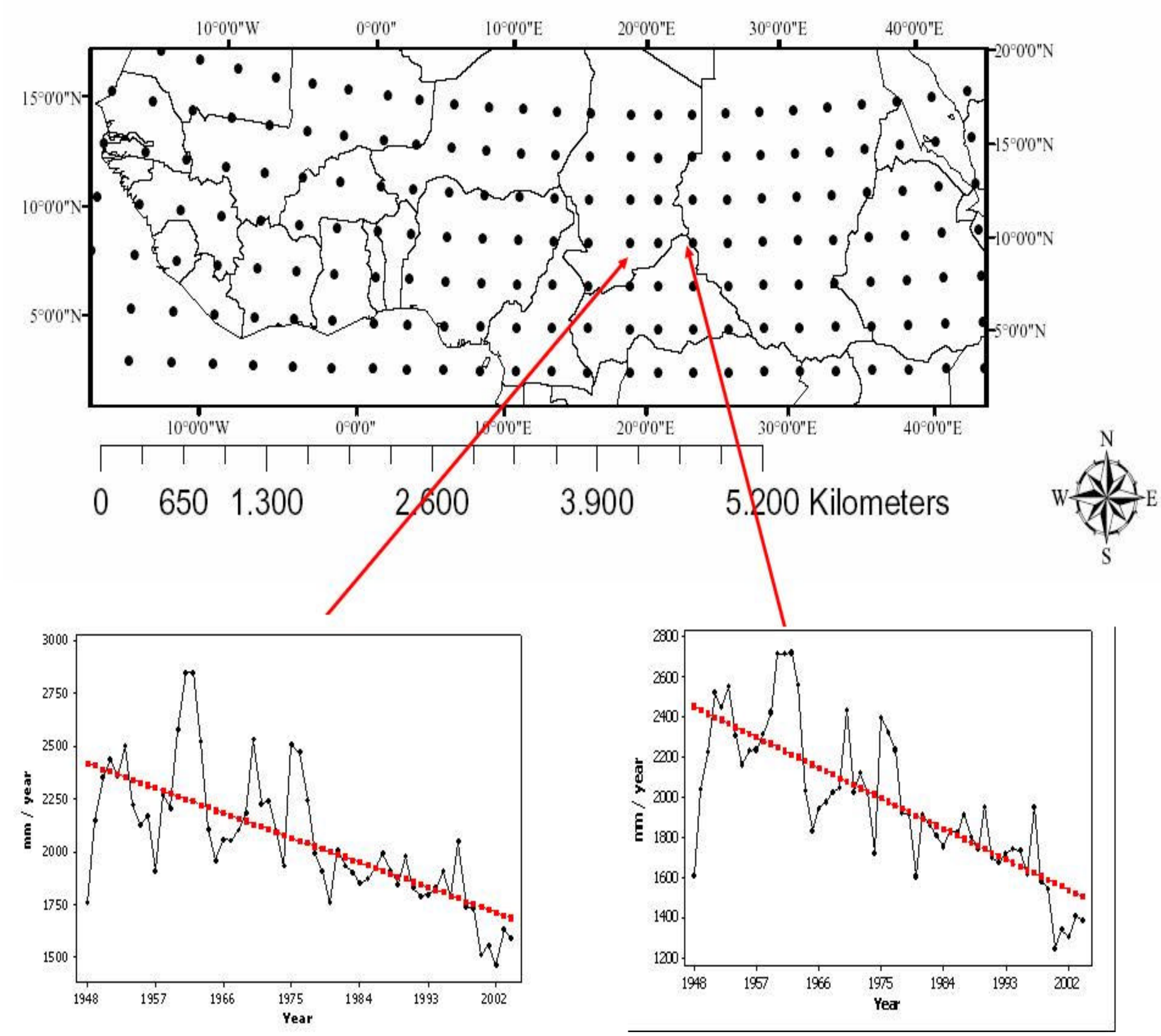


Figure 28: Precipitation trends of two grid point averages in East Africa (10.5 º N 30.75 º E and 10.5 º N 25.75 º E)

Plot of the data from three climatic periods 1951-1980, 1961-1990, and 1971-2000 are shown in Figures 29, 30, 31. The average precipitation trend for the first period 1951 to 1980 was 140 mm per decade, for the second period 203 mm per decade, and the trend for the third period 1971-2000 was 49 mm per decade. The results show fluctuations in precipitation change over the 3 climate periods.

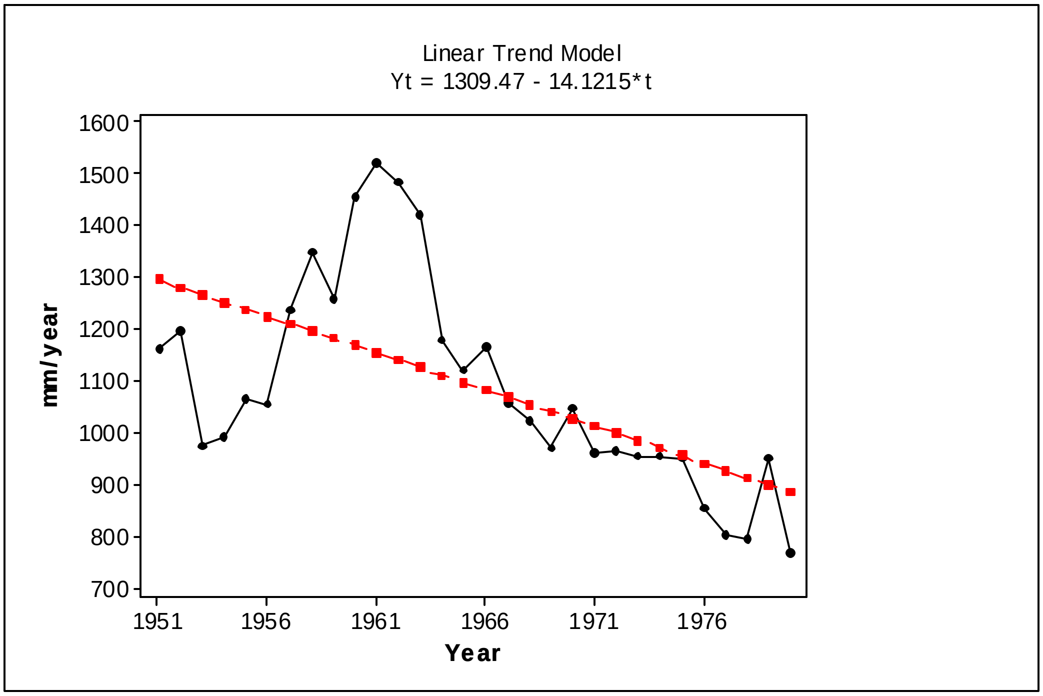


Figure 29: Annual mean precipitation trend 1951-1980.

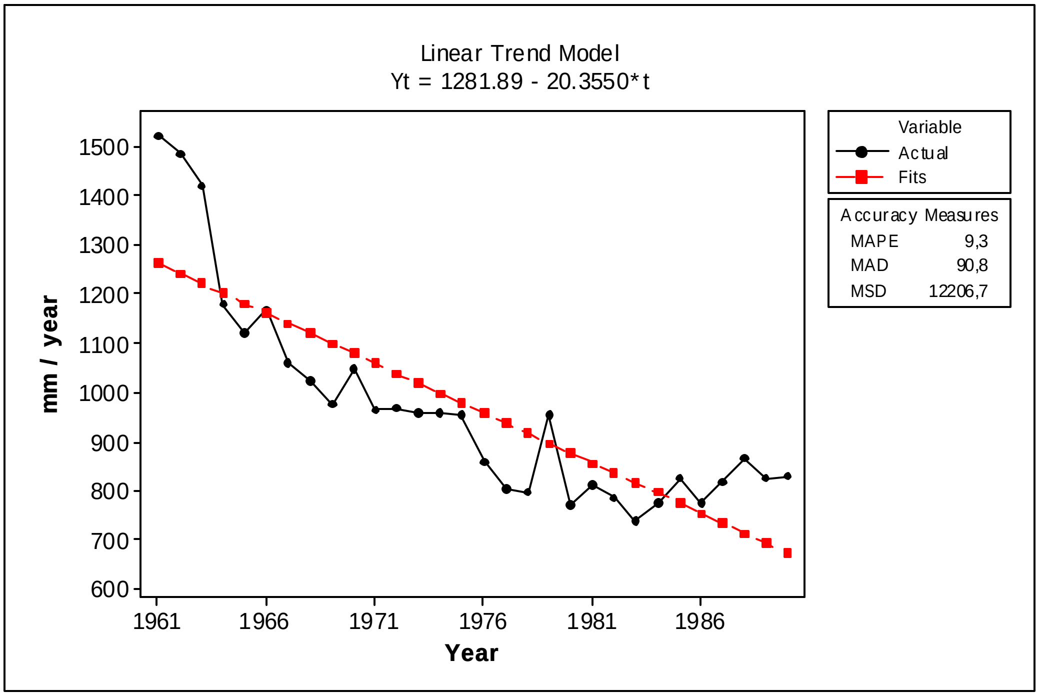


Figure 30: Annual mean precipitation trend 1961-1990.

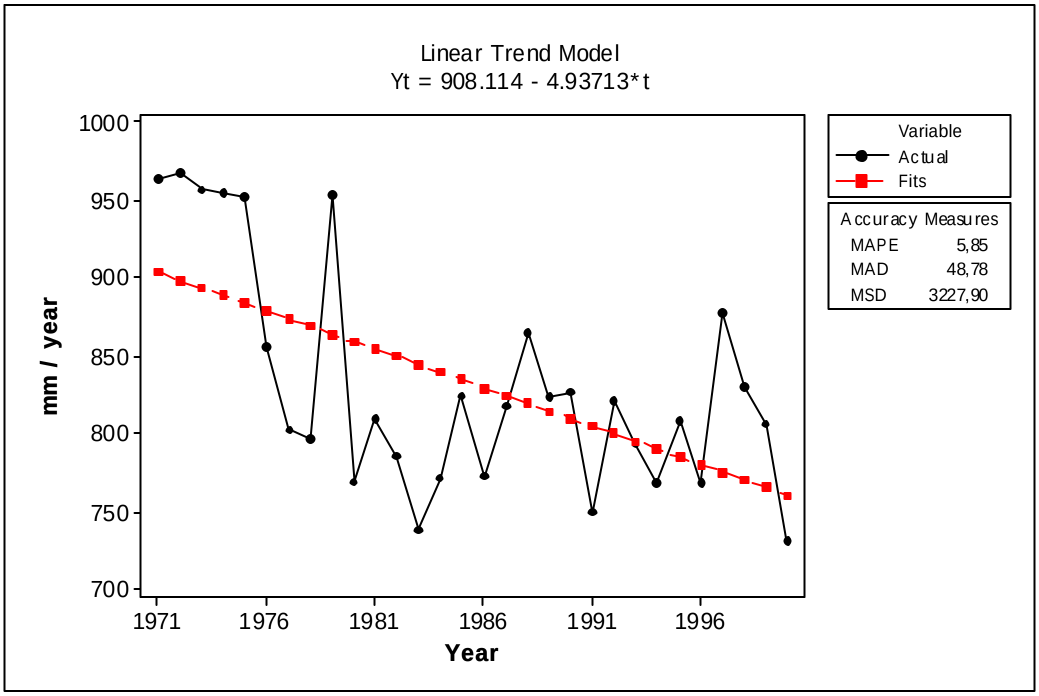


Figure 31: Annual mean precipitation trend 1971-2000.

Figures 32, 33 present precipitation trends for June – September and March – May raining seasons in the Sahel and East Africa. The distribution of precipitation data for the Sahel is 100 – 600 mm / year and seasonal precipitation has increased at an estimated rate of 4 mm /year. In East Africa precipitation is distributed between 400 – 1800 mm/ year and seasonal precipitation has decreased at a rate of -20 mm / year. However these results are not statistically significant as significance testing provided P values of 0.2 and 0.17 for the trends from Sahel and East Africa respectively.

Results presented in Table 7 show that for the Sahel, 1988-1997 recorded the highest average seasonal precipitation / decade with a mean of 486 mm / decade and estimates for 1958-67 and 1978- 87 were significantly lower than the average mean for the entire Sahel, 279 mm/ decade. In East Africa the lowest average precipitation average / decade was in 1998- 2004, average decadal precipitation decreased sturdily from1958-67, till it reached a minimum in 1998 -2004.

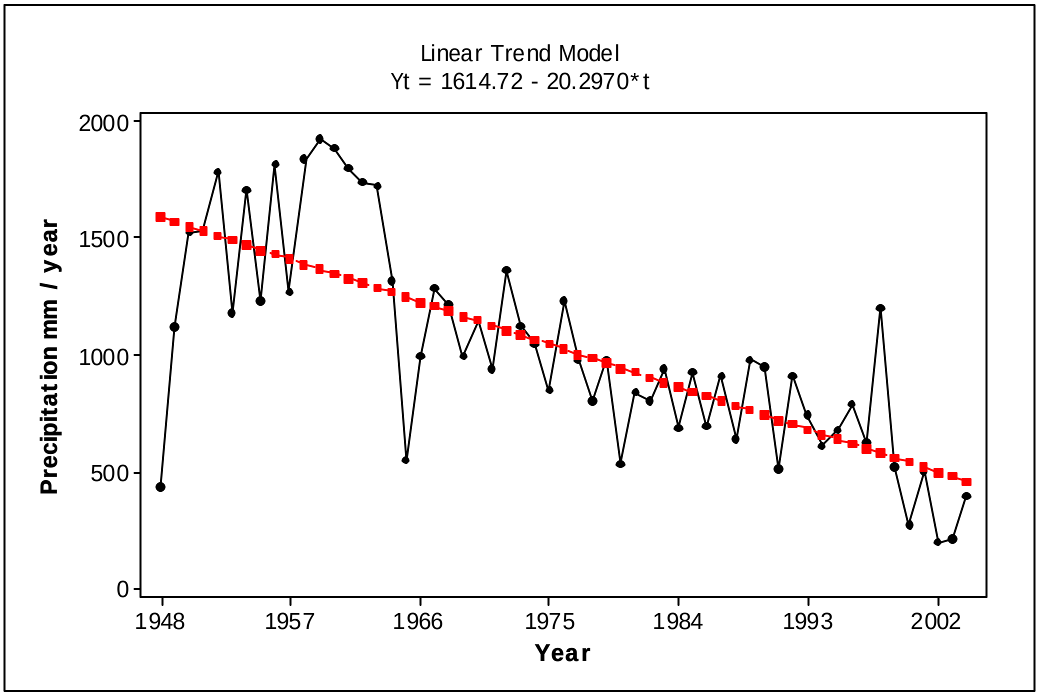


Figure 32: Annual mean precipitation trend for March – May raining season, East Africa

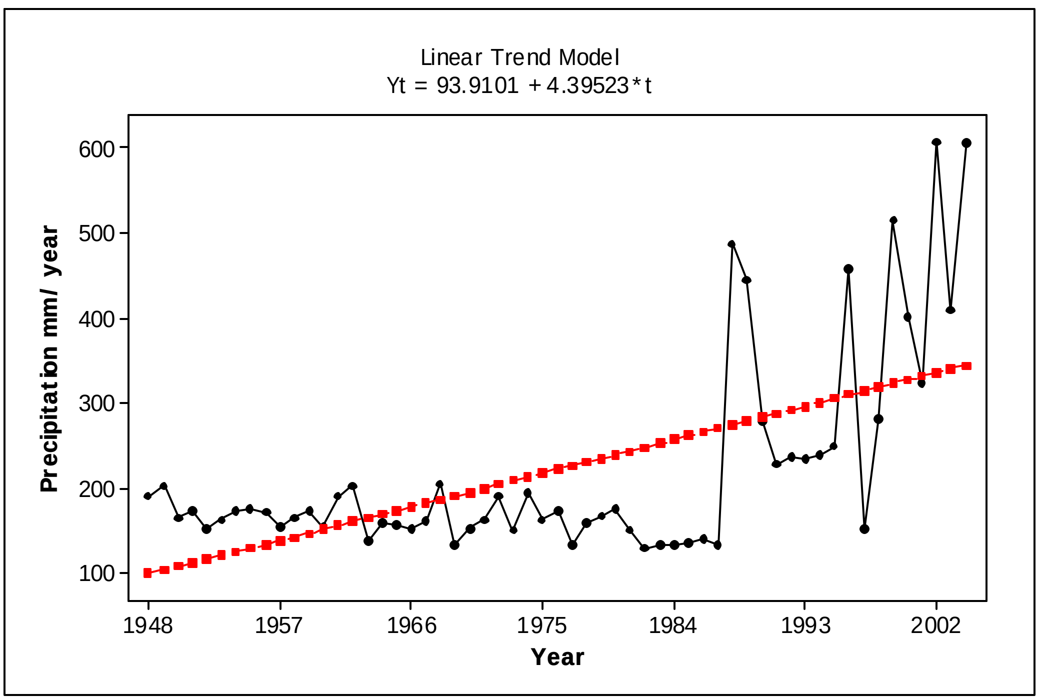


Figure 33: Annual mean precipitation trend for June – September raining season, Sahel

Table 7: Table of decadal precipitation means (June to September; March – May) raining seasons from 1948 through 2004, and the 57-year mean for Sahel and East Africa

Precipitation mm/decade							
	1948-57	1958-67	1968-77	1978-87	1988-97	1998-2004	57 year mean
Sahel	189	169	204	158	486	477	279
East Africa	454	1841	1218	807	640	349	885

4.3 Climate change in Central Africa

Results presented in sections 4.1 and 4.2 shows a significant decrease in precipitation and increase in temperature over Central Africa for the period 1948 - 2004. The results reveal a change in climate regime beginning in the early 1970s with the last three decades of the 20th century presenting a dry episode. Previous researchers e.g. Kidson (1977), Rognon (1987), Nicholson (1978) have also mentioned this episode existing over Central Africa. In order to understand its climatic significance, it is necessary to address past Central African rainfall variability on longer timescales. The crucial question is whether or not this period is unique within the epoch during which a climatic regime similar to that of the twentieth century has prevailed in Central Africa. If historical precedents for the recent desiccation exist, the likelihood that it is a precursor of indefinite enhanced aridity is reduced, although not negligible. Even if the recent desiccation is unique (or at least highly unusual), it does not necessarily presage a drier Central Africa in the near future, but may suggest that there has been a change in the mechanisms that modulate Central African rainfall which enhances the likelihood of protracted dry conditions. Such changes may arise from shifts in the regional or global circulation, as is suggested by some of the research discussed in Chapter 2. Alternatively, a reduction in rainfall may be associated with changes in local or regional-scale feedback mechanisms.

The historical context of the recent dry episode may be assessed through investigations of past climatic variations as indicated by long term climate data (century timescale). This approach is limited by the availability and nature of climate records in Central Africa. Such information is generally of limited temporal and spatial coverage, and does not allow detailed, quantitative climatic reconstructions to be generated at a high enough temporal resolution for the purposes outlined above. Nonetheless, such information should be incorporated into any plausible model of long-term climatic variation where possible. Such data as exist may be used to assess and validate representations of the regional climate in long-term climate simulations generated by global climate models (GCMs). If such simulations are capable of reproducing known climatic variations, they may be used to investigate the mechanisms associated with changes in rainfall such as those observed since the early 1970s. Where GCM simulations under-perform in representing regional climate variations, a consideration of the differences between the simulation and reality in terms of the processes included in, and omitted from, the model parameterisations may also indicate the relative importance of certain processes in determining the regional climate.

It is known from historical sources (Nicholson, 1976) and palaeoclimate studies (Butzer, 1983; Holmes et al., 1997) that Central Africa has experienced periods of drought and desiccation in the past, recurring on timescales of decades to centuries. What is uncertain, however, is the duration and intensity of these dry episodes. An investigation into the frequency and distribution of such dry periods over the geologically recent past should further our understanding as to whether the recent desiccation is

(i) a precursor of indefinite enhanced aridity in the region

(ii) representative of a regular, if infrequent, phenomenon of drought lasting for several decades,

(iii) simply an unusually long, but temporary, dry episode of the type familiar from the historical record.

The answer to this question has important implications for hypothesised causes of Central African aridity occurring on decadal timescales (Street-Perrott and Perrott, 1990; Hulme and Kelly, 1993; Rowell et al., 1995). From the point of view of placing the recent Central African dry episode in an historical context, we require a knowledge of rainfall variations over a period during which the regional (and, arguably, the global) climatic regime has been broadly similar to that of the present day. It is therefore pertinent to examine climatic variability in the region since the onset of the Saharan desiccation which occurred some 5000 years ago (Lioubimtseva, 1995; Jolly et al., 1998).

Apart from revealing a drying trend in the last three decades over Central Africa, results also shows different responses of local climate to regional climate change. Results provided in maps in Figures 17 and 26 show two main areas with different responses in respect to change in climate over Central Africa. The eastern part of Central Africa is characterized by a steep drying trend with marked precipitation decline and the other area by an increase in precipitation and temperature decline (Sahel).

The impact of climate change is being faced every single year by peasant farmers in the eastern parts of Central Africa particularly in countries such as Ethiopia, Sudan and Somalia. In recent years, rainfall that used to start as early as March now does not start until June and then stops in August instead of October. Results in Figure 32 show that the raining seasons have shrunk and decreased at a rate of -20 mm/ year and there simply is not enough time to grow the crops. For some of Africa's poorest countries in which the majority of our 74 million people rely on rain, this spells disaster. The truth is that climate change has affected people's ability to grow crops, rear livestock and find water to drink. Even malaria has become more and more life threatening in many places as it spreads to warming regions. Currently the number of studies going on in these areas to determine the cause of the drying trend is very limited and this study recommends further studies in the eastern parts of Central Africa to determine the cause of the excessive drying in this area.

The effects of the severe drought in 1983–1984, affecting certain parts of Central Africa such as the Sahel were documented using satellite imagery by several authors, notably Tucker et al. (1991). They showed a marked southward shift of the edge of the desert, but noted that it would require long-term observations to prove a secular change, given the notoriously variable climate of the area. More recent studies in the Sahel have suggested secular trends of change, many related to vegetation and land use (Mortimore and Adams, 2001; Reij and Thiombiano, 2003), at a national scale (Niemeijer and Mazzucato, 2002) and even at the regional to subcontinental scale (Eklundh and Olsson, 2003). The greening of the desert which scientists say has been happening largely unnoticed since the mid-1980s contradicts the commonly held idea that desertification is an irreversible process. Results presented in this study in Figures 26 confirm the increase in precipitation in the Sahel as recent studies have suggested. What is still unclear is the cause of the greening; increasing rainfall over the last few years is certainly one reason, but does not fully explain the change. Other factors, such as land use change and temperature, may also contribute. Section 4.4 discusses the greening of the Sahel and examines possible causes.

4.4 Vegetation Change in the Sahel

"The Sahel," as described by Anyamba and Tucker (2005), "is a semi-arid region stretching approximately 5000 km across northern Africa from the Atlantic Ocean in the west to near the Red Sea in the east and extending roughly from 12°N to 18°N," which "forms an ecological transition between the Sahara desert to the north and the humid tropical savannah to the south (Le Houerou, 1980)." Researchers studying the Sahel today focus on the continued economic fragility, its halting steps towards democratic political regimes, and its continued food security problems. Despite complex economic migration patterns and urban expansion in the 20th century, the vast majority of the rural dwellers are dependent on some form of rain-fed agriculture or animal production. Some suggest that there are no "normal" rainfall levels in this region; just fluctuating supplies and changing human demand for water. Three major droughts have occurred this century, in 1910-1916, 1941-1945, and a long period of below average rainfall (termed 'desiccation') that began in the early-1970s and continued, with some interruptions, into the 1980s.

The hazardous conditions of the droughts of the 1970s, and those that followed, have had cumulative impacts, but these impacts form part of complex patterns of social and economic change, and it is almost impossible to separate the effects of the natural hazard (drought) from other factors that made individuals vulnerable. Vulnerability is an everyday situation for some people, but a rare occurrence for others. It is important here to differentiate between meteorological drought - below-average moisture supply - and the effects of changing human land uses and practices. Low rainfall can be coped with, if farmers have a diverse livelihood system, or sufficient assets. Famine situations have resulted in dryland West Africa where drought conditions have surprised the population that were unprepared for it (as in the 1970s, when fifteen years of good rainfall had encouraged many to over-invest in agriculture). The possible range of adjustments has been constrained by warfare, social status, or corruption and mismanagement. In some areas people starved without drought conditions, because of locust invasions, epidemics, or the seizure of their harvests by warlords or even colonial administrators.

Climate change in the Sahel challenges researchers and policy makers because the best models around predict opposite outcomes (Brahic, 2006). Computer models of the future climate often disagree about the scale of likely change but predictions for the Sahel are also contradictory about the direction of change and give policy makers little help in preparing for the future.

Working with National Oceanic and Atmospheric Administration (NOAA) Advanced Very High Resolution Radiometer (AVHRR) data obtained from polar orbiting satellites, Anyamba and Tucker developed a Normalized Difference Vegetation Index (NDVI) history that stretches from 1981 to 2003. Comparing this history with the precipitation history of the Sahel developed by Nicholson (2005), they find that "the persistence and spatial coherence of drought conditions during the 1980s is well represented by the NDVI anomaly patterns and corresponds with the documented rainfall anomalies across the region during this time period." Thereafter, they also find that "the prevalence of greener than normal conditions during the 1990s to 2003 follows a similar increase in rainfall over the region during the last decade."

Olsson et al. (2005) , in another analysis of NDVI and rainfall data finds "a consistent trend of increasing vegetation greenness in much of the Sahel," which they describe as "remarkable," and state that increasing rainfall over the last few years "is certainly one reason" for the greening phenomenon. However, they find that the increase in rainfall "does not fully explain" it.

Only eight out of 40 rainfall observations showed a statistically significant (95%) increase of rainfall between 1982-1990 and 1991-1999. In addition, they report that "further analysis of this relationship does not indicate an overall relationship between rainfall increase and vegetation trend." Amongst the factors affecting the growth of plants particularly during germination, growth and during photosynthesis is temperature (thermoperiod). Thermoperiod refers to the daily temperature change. Plants produce maximum growth when exposed to a day temperature that is about 5.5 to 8°C higher than the night temperature. This allows the plant to photosynthesize (build up) and respire (break down) during an optimum daytime temperature, and to curtail the rate of respiration during a cooler night. High temperatures cause increased respiration, sometimes above the rate of photosynthesis. This means that the products of photosynthesis are being used more rapidly than they are being produced. For growth to occur, photosynthesis must be greater than respiration.

Low temperatures can result in poor growth because photosynthesis is slowed down. As a consequence, growth is also slowed, and this results in lower yields. Not all plants grow best in the same temperature range. For example, snapdragons grow best when night time temperatures are 13 °C, while the poinsettia grows best at 18 °C. Florist cyclamen does well under very cool conditions, while many bedding plants grow best at higher temperatures.

Buds of many plants require exposure to a certain number of days below a critical temperature (chilling hours) before they will resume growth in the spring. Peaches are a prime example; most cultivars require 700 to 1,000 hours below 45°F and above 32°F before they break their rest period and begin growth. This time period varies for different plants. The flower buds of forsythia require a relatively short rest period and will grow at the first sign of warm weather. During dormancy, buds can withstand very low temperatures, but after the rest period is satisfied, buds become more susceptible to weather conditions, and can be damaged easily by cold temperatures or frost.

Results from this study show a decline of temperature of -0.06 to -0.02 K per year in the Sahel. This decline started in the early 1970s and may be one of the factors influencing the greening of the Sahel. So far explanations for the greening of the Sahel have neglected the influence of temperature and this study argues that temperature decline amongst other factors may affect the thermoperiod of plants within the Sahel region resulting in their ultimate growth. The study proposes a further study on the vegetative and optimum thermoperiods of plants in the Sahel in order to prove or disprove this fact.

Olsson et al. (2005) suggest that another potential explanation could be improved land management, which has been shown to cause similar changes in vegetation response elsewhere (Runnstrom, 2003). However, in more detailed analyses in Mali, where production of millet rose by 55% and 35%, respectively, since 1980, they could find "no clear relationship" between agricultural productivity and NDVI, which argues against the land management explanation.

A third speculation of Olsson et al. (2005) is that the greening of the Sahel could be caused by increasing rural-to-urban migration. In this scenario, widespread increases in vegetation occur as a result of "reduced area under cultivation," due to a shortage of rural labourers, and/or "increasing inputs on cropland," such as seeds, machinery and fertilizers made possible by an increase in money sent home to rural households by family members working in cities. However, Olsson et al. (2005) note that "more empirical research is needed to verify this [hypothesis]."

Since the mid 1980s several computer models have suggested that changes in the surface temperature of the oceans have changed the dynamics of the West African monsoon and are therefore to blame. The hypothesis gained widespread support but there is still some disagreement. Different models point the finger at different oceans – some say the influence of the Indian oceans is most important, others the difference between the North and South Atlantic.

Most scientists agree that the greenhouse gases and aerosols that human activities released into the atmosphere are partially to blame for changing ocean temperatures. The question is how this will affect future rainfall? Again, the answers depend on the model used. Hoerlings (2006) looked at all the most recent climate models, averaged them out and came to the conclusion that the Sahel`s recent fate would be reversed in the 21st Century. Global warming, Hoerlings (2006) concluded, would bring much needed rainfall to the region, one of the very few positives outcomes of greenhouse gas emissions.

But Held (2005) published results of a new climate model that suggested that, far from becoming wetter, the Sahel faces a period of dynamic drying if green house gases are not checked. He admits that his results, from a single model, should not be the basis for policy decisions. But one striking point meant the results could not be discarded. The model mimicked the regions recent climate more faithfully than any previous one had – an important measure of how reliable it is.

Understanding why the models predict such widely divergent futures is a significant priority that requires really getting into the bowels of the models (Giannini, 2006). There must be something in the models physics that is causing them to respond differently. For instance, several researchers have pointed out that many of the models show cooler present day sea temperatures near the Americas and warmer ones close to Africa while the reality is the other way around suggesting that the models are flawed.

Hurrell (2006) attributes the difficulties in modelling future Sahel rainfall to the multiple competing influences of factors that have comparable importance. Hurrell (2006) suggests that global sea surface temperatures play a strong and possibly dominating role in determining how much rain falls in the Sahel- more so than for instance, temperatures above Africa. But the problem gets more difficult as evidence points to the Atlantic, Indian and Pacific Ocean all playing a role through different mechanisms. The different parts of the Sahel might be affected differently by the relative influence of each ocean. To complicate matters, the relative importance of factors affecting the Sahel`s climate is tipped in different directions by different models.

4.5 Drying in Eastern Central Africa

Overall Africa has warmed 0.7 K over the 20th century and general circulation models project warming across Africa ranging from 0.2 K per decade (low scenario) to more than 0.5 K per decade (high scenario), see Hulme et al. (2001); IPCC (2001). Results from this study show a rise in temperature of 0.15 K per decade and a decrease in precipitation of -91-mm per decade over Central Africa. The decrease in precipitation is severely marked in the eastern part of the study area particularly in countries such as

Ethiopia, Sudan, and Somalia. Precipitation patterns in the eastern part of Central Africa are more variable.

Projections of climate change suggest that Eastern Central Africa will experience warmer temperatures and a 5-20% increased rainfall from December-February and 5-10% decreased rainfall from June-August by 2050 (Hulme et al. 2001; IPCC, 2001). Not only are these changes not uniform throughout the year, they occur in sporadic and unpredictable events. It is also expected that there will be less precipitation in East Africa during the already dry droughts and increased desertification in the region.

Recent research also suggests that warming sea surface temperatures, especially in the southwest Indian Ocean, in addition to inter-annual climate variability (i.e., El Niño/Southern Oscillation (ENSO)) may play a key role in East African rainfall and may be linked to the change in rainfall across some parts of equatorial-subtropical East Africa (Cane et al., 1986; Plisnier et al., 2000; Rowe, 2001). Warm sea surface temperatures are thought to be responsible for the recent droughts in equatorial and subtropical Eastern Central Africa during the 1980s to the 2000s (Funk et al., 2005).

However because temperature has increased and precipitation in the region has decreased in some areas, Eastern Central Africa is already affected. For example, from 1996 to 2003, there has been a decline in rainfall of 50-150 mm per season (March to May) and corresponding decline in long-cycle crops (e.g., slowly maturing varieties of sorghum and maize) across most of eastern Africa (Funk et al., 2005). Long-cycle crops depend upon rain during this typically wet season and progressive moisture deficit results in low crop yields in the fall, thereby impacting the available food supply.

Inter-annual climate variability (e.g., ENSO) has huge impacts on the region's climate. Warm ENSO events also referred to as El Niño events produce abnormally high amounts of precipitation in parts of equatorial Eastern Central Africa and can result in flooding and decreased agricultural yields. Warming temperatures are projected to cause more frequent and more intense extreme weather events, such as heavy rain storms, flooding, fires, hurricanes, tropical storms and El Niño events (IPCC, 2001). Tropical storms can ravage coastal areas and intensive the impacts of sea-level rise by accelerating erosion in coastal areas and by removing protective natural buffer areas that absorb storm energy, such as wetlands and mangroves (Magadza, 2000). Extreme rainfall and subsequent heavy flooding damage will also have serious effects on agriculture including the erosion of topsoil, inundation of previously arid soils, and leaching nutrients from the soil. Regional fluctuations in lake levels are another impact of regional climate variations and are expected to worsen with projected climate change. While land use change can have a dramatic effect on lake levels, climate variability is more unpredictable and difficult to manage for. For example, lake levels in Lake Tumba in the Democratic Republic of Congo (Inogwabini et al., 2006) and Lake Victoria in Kenya (Birkett et al., 1999; Latif et al., 1999) have been attributed to climate variations and may become more variable in the future.

In 1997, floods and high rainfall, triggered by an El Niño event in eastern Africa, resulted in a surface rise of 1.7 meters in Lake Victoria and disrupted agricultural production and pastoral systems (Lovett et al., 2005). While climate change is projected to cause more frequent and intense ENSO events (Wara et al., 2005), impacts are not uniform across East Africa. In fact, the same year that the waters were rising in Lake Victoria, El Niño triggered a severe drought in another location in Kenya, significantly decreasing hydro-electric power output and thus limiting the availability of electricity to East Africans (Lovett et al., 2005). Further, a projected increase in precipitation may also have an effect on hurricanes and storms in the Atlantic. Landsea and Gray (1992) have found that rainfall in the Sahel is positively correlated with the intensity of hurricanes in the Atlantic Ocean.

4.6 Spatio-temporal patterns of Central African climate

Figure 34 is a summary of the PCA analysis for extracting the dominant spatio temporal modes of annual rainfall variability, compressing the original $224\ (\text{grid points}) \times 684\ (\text{number of points})$ data matrix into a 4×684 data matrix. Four patterns stand out, selected based on the scree plot Figure 35; the four patterns combined capture 65.3 % of the total variance in the precipitation over Central Africa. The selected principal components explain 37.6%, 12.1%, 10.7% and 4.9% respectively of the total variance in the precipitation dataset. Table 8 presents the selected principal components, their corresponding eigenvalues and percentage variance.

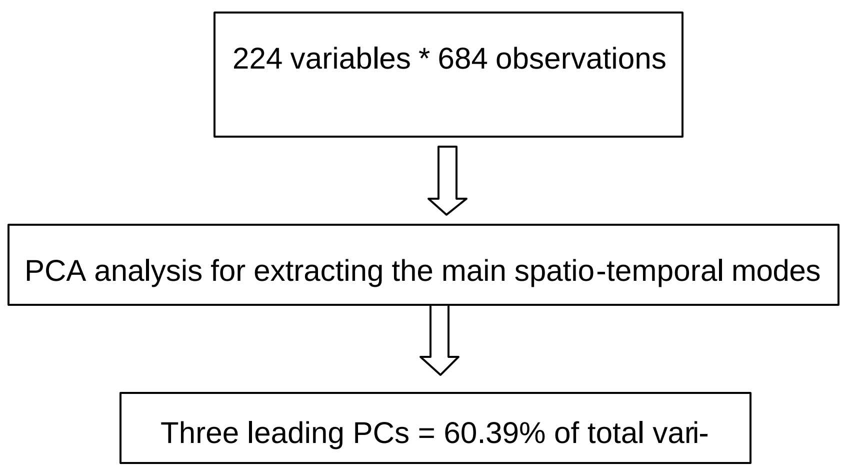


Figure 34: Scheme describing the steps used in PCA analysis

Table 8: Table showing the eigenvalues and percentage variance of principal components from analysis of precipitation dataset

	PC 1	**PC 2**	**PC 3**	**PC 4**
Eigenvalue	84.18	27.09	23.98	**11.01**
% variance	37.6	12.1	10.7	**4.9**
% cumulative	37.6	49.7	60.4	65.3

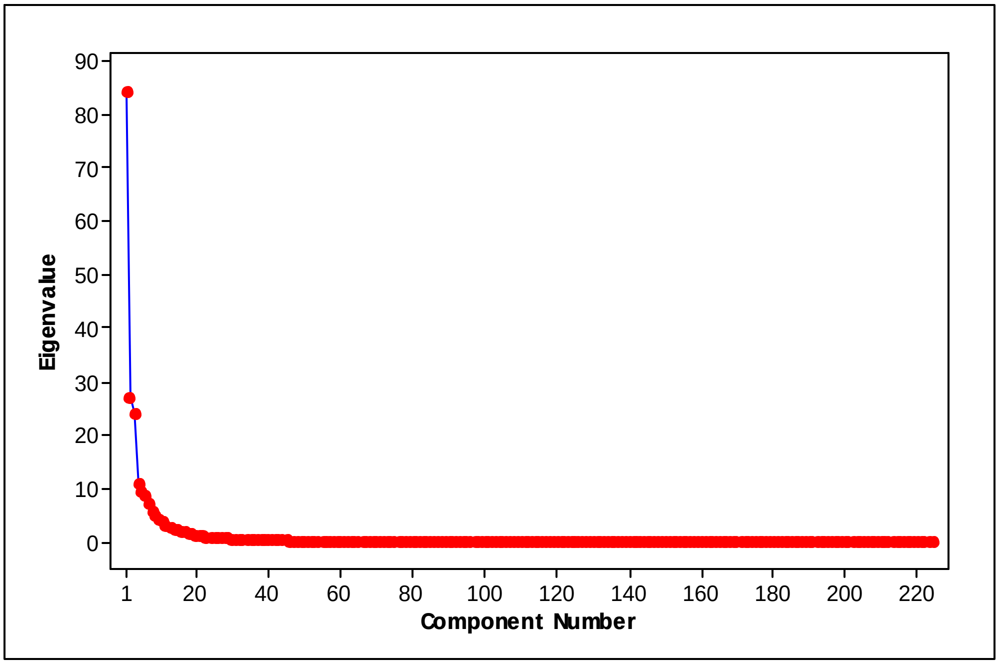
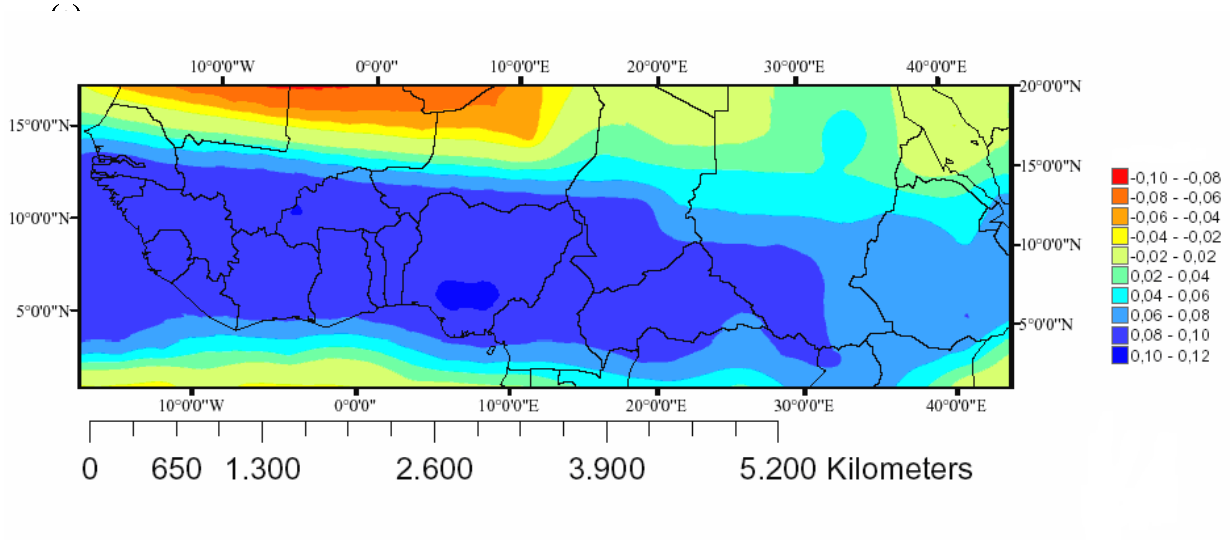


Figure 35: Scree plot from principal component analysis of precipitation dataset.

In one of the first continental scale studies in Africa Nicholson (1986) identified two recurrent spatial patterns of precipitation over Africa. One was characterized by continent wide anomalies of the same sign, the other by anomalies of opposite sign at equatorial latitude and at the poleward margins of the monsoon region. Nicholson's findings are echoed in this analysis, however on the regional scale. The leading pattern (PC 1) of annual mean precipitation from the period 1948-2004 is illustrated in Figure 36a. It takes on stronger positive values in the mid-western part of the study area particularly along the coastal Gulf of Guinea and progressively weaker but significant positive values to the mid-eastern (Sudan, Somalia, Ethiopia) part of the study area. To the north, the values of the spatial pattern of the first PC tend to be progressively negative. This pattern is consistent with the first eigenvector pattern for Africa analyzed by Giannini et al. (2006). Results from the study of Giannini et al. (2006) revealed two distinct patterns of precipitation over Africa. One in the northern hemisphere characterized by higher positive values which includes along the Coast of Gulf of Guinea and the other in the southern hemisphere which is characterized by weaker values. The patterns from the first PC combine all features expressed in the northern and southern hemisphere as revealed by the work of Giannini et al. (2006). The linear trend of the associated time series for the first principal component presents is overall decreasing (Figure 36b). Fluctuations with a multi-decal time scale superimposed on year to year variations are evident in the associated time series of the first principal component provided in Figure 36b. The 1950s and 1960s were clearly wetter than the long term average. The wetter than average central decades of the 20th century certainly contribute to making the contrast with the dryer decades in the 1970s and 1980s, even larger. This pattern is statistically related to global SSTs, with the sign such that drying over Africa is associated with a warmer tropical pacific, Indian and south Atlantic oceans and a cooler north Atlantic basin (Giannini et al. 2006). The spatial and temporal patterns of the first principal component are actually representative of regional anomalies demonstrated by a significant correlation of the associated time series of the first principal component with the regional averages described in the previous subsection (over 1948-2004); the Central African average of annual precipitation is unequivocally correlated with the drying continent pattern (correlation 0.979). Previous studies have proposed two explanations for this drying pattern. These include:

a) Moisture supply argument as suggested by Lamb (1978) and Hoeling et al. (2006). They emphasize the role of the north-south SST gradient which is evident in the Atlantic basin. This explanation relates to the recent drying of parts of Central Africa notably the Sahel to an equator-ward shift of the mean location of the Atlantic ITCZ, connecting reduced oceanic precipitation with a weaker West African monsoon flow.

b) A second explanation consistent with modelling results of Gianinni et al. (2003) and Lu and Delworth (2005) is a stabilization argument. The argument suggested that warmer Indo-Pacific SSTs causes warming of the entire tropical troposphere, stabilizing the tropical troposphere from above in the Atlantic sector and over Africa. The key to this mechanism is that the continental boundary layer is unable to increase its energy content and buoyancy to keep the energy content over the surrounding oceans. The issue of if Central African rainfall is affected not only by ocean temperature gradients, but also by spatially uniform increases in oceanic temperatures, is central to projections of climate change in the 21st Century, Gianinni et al. (2006).



(b)

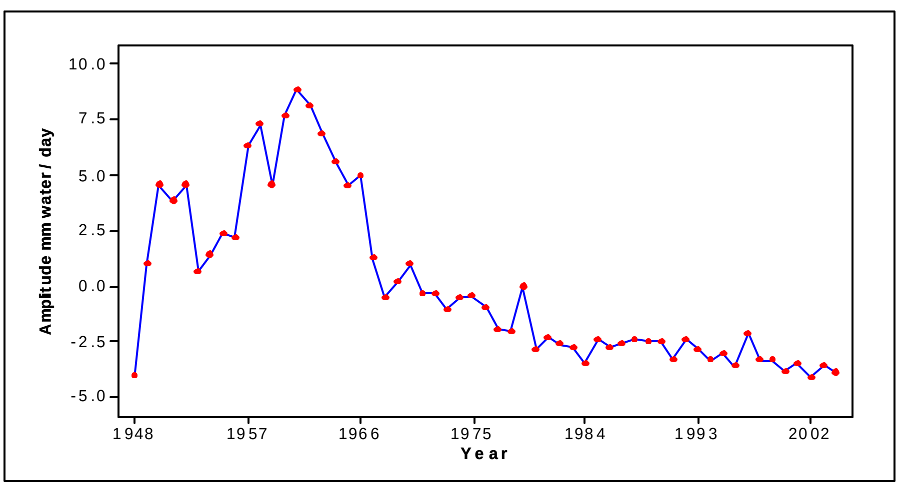
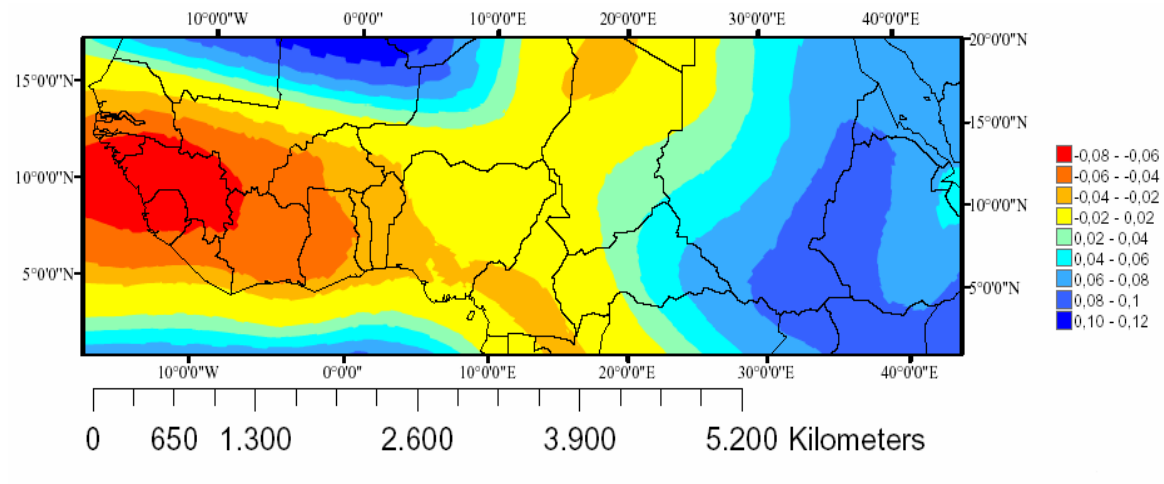


Figure 36: (a) Spatial loading of the first principal component from PCA analysis of precipitation over Central Africa (b) corresponding time series of PC1

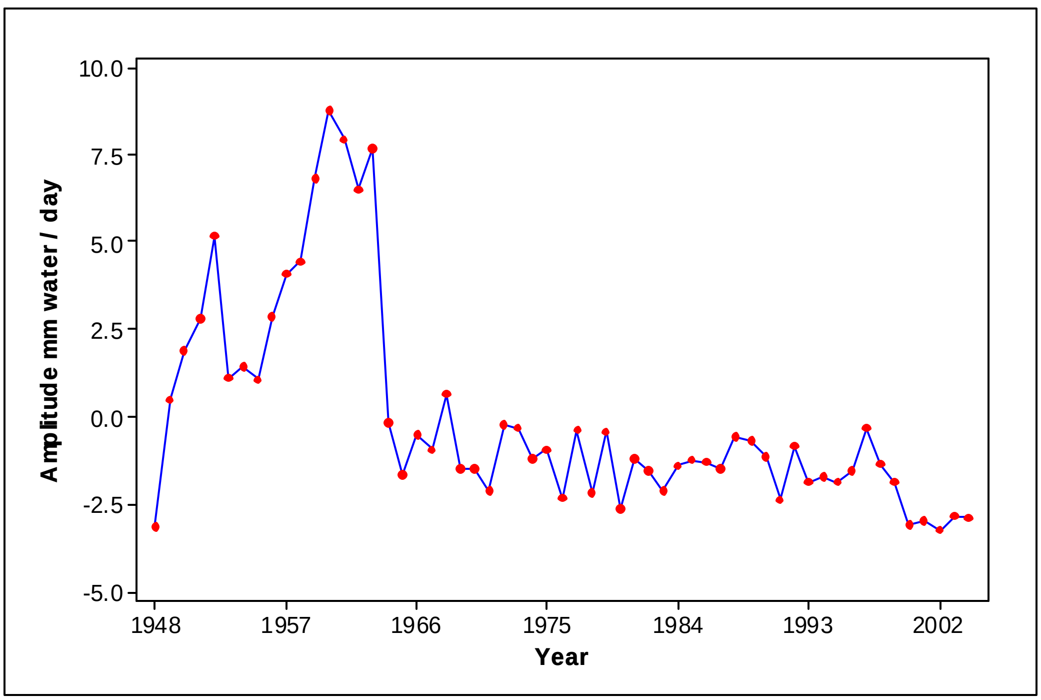
The illustrated patterns of the second principal component, Figure 37a, which explains 12.1 % of the variance in the precipitation dataset exhibits a dipole pattern over Central Africa with positive loadings in the eastern and negative loadings in the western parts. Fluctuations with multidecal time scale superimposed on year to year variations are evident in the time series of the associated factor score of the second principal component, Figure 37b. The linear trend of the associated time series of the second principal component is decreasing revealing drying in Central Africa. However a further examination of the associated factor score of the second principal component on a monthly scale reveals two peaks in the monthly time series, one in April and the other in November Figure 37c.

Together the spatial and temporal patterns of the second principal component of precipitation over Central Africa capture the different seasons (dry and wet seasons) in the different climatic regions within Central Africa. Seasons over Central Africa result from interactions between atmosphere, land and ocean as they respond to the annual cycle of insolation. In Central Africa, heavy rains occur in zonally symmetric, meridionally confined bands or rain belts which straddle the equator near the equinoxes in April and October and move northward in July and Southwards in January. The peaks in April and November in the monthly time series of the associated factor scores depict the wet season in some parts of Central Africa particularly in the Eastern parts and the lows between June and September depict dry seasons over other parts of Central Africa particularly in the western parts and in areas such as the Sahel.

(a)


10°0'0"W
0°0'0"
10°0'0"E
20°0'0"E
30°0'0"E
40°0'0"E
20°0'0"N
15°0'0"N
10°0'0"N
5°0'0"N
-0,08 - -0,06
-0,06 - -0,04
-0,04 - -0,02
-0,02 - 0,02
0,02 - 0,04
0,04 - 0,06
0,06 - 0,08
0,08 - 0,1
0,10 - 0,12
0 650 1.300 2.600 3.900 5.200 Kilometers

(b)


Amplitude mm water / day
10.0
7.5
5.0
2.5
0.0
-2.5
-5.0
1948
1957
1966
1975
1984
1993
2002
Year

(c)

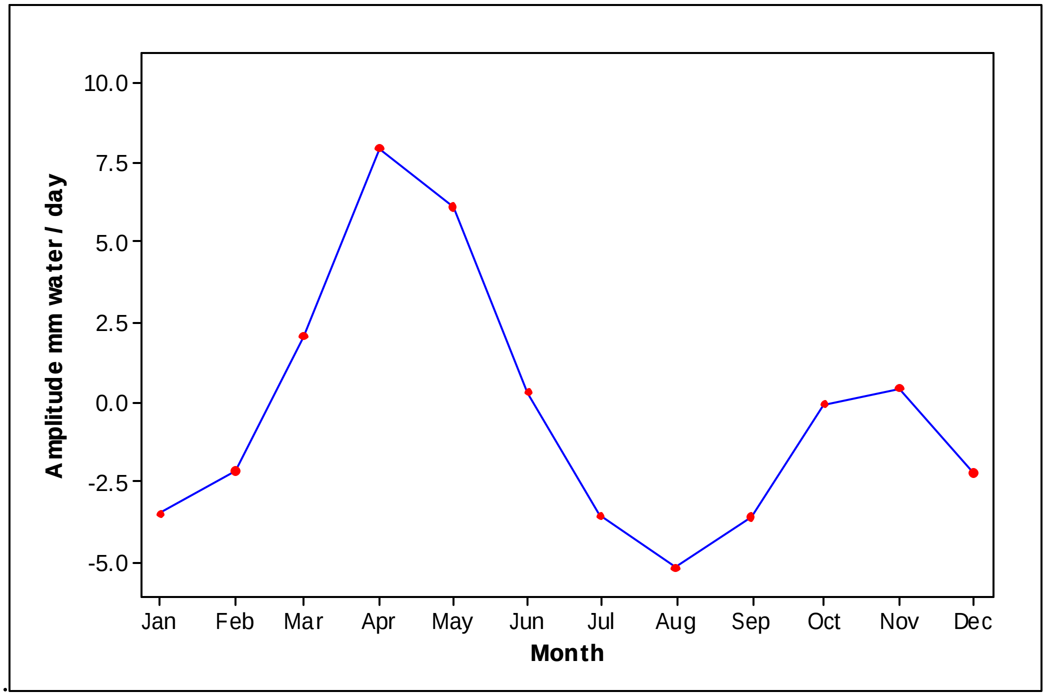
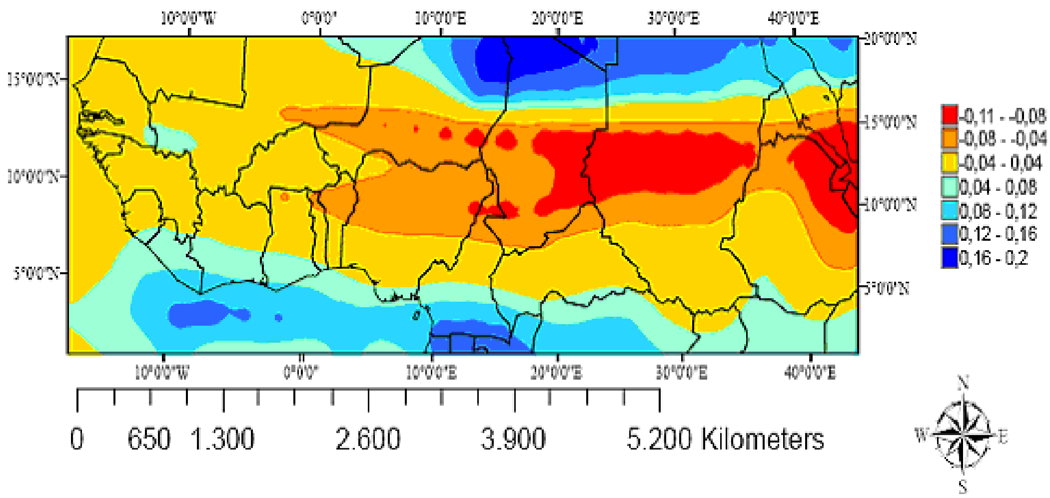


Figure 37: (a) Spatial loading of the second principal component from PCA analysis of precipitation over Central Africa (b) corresponding annual time series of factor scores of PC2 (c) monthly time series of the factor scores of PC2

The third principal component represents 10.7% of the variance. Figure 38a show its spatial eigenvector pattern. Negative or weakly positive loadings are observed from the western through to the mid eastern parts of the study area. The mid northern parts have positive loadings. The time series of the factor scores of the third principal component is dominated by strong year to year variability. The time series shows two clear periods one dominated by negative loadings (1948-1966) and the other dominated by positive loadings with the change occurring in the late 1960s. The spatial and temporal loadings of the third principal component represent the change in climate regimes over Central Africa which occurred in the early 1970s. This change in precipitation regime has resulted in negative loadings in most parts of Central Africa.

(a)



(b)

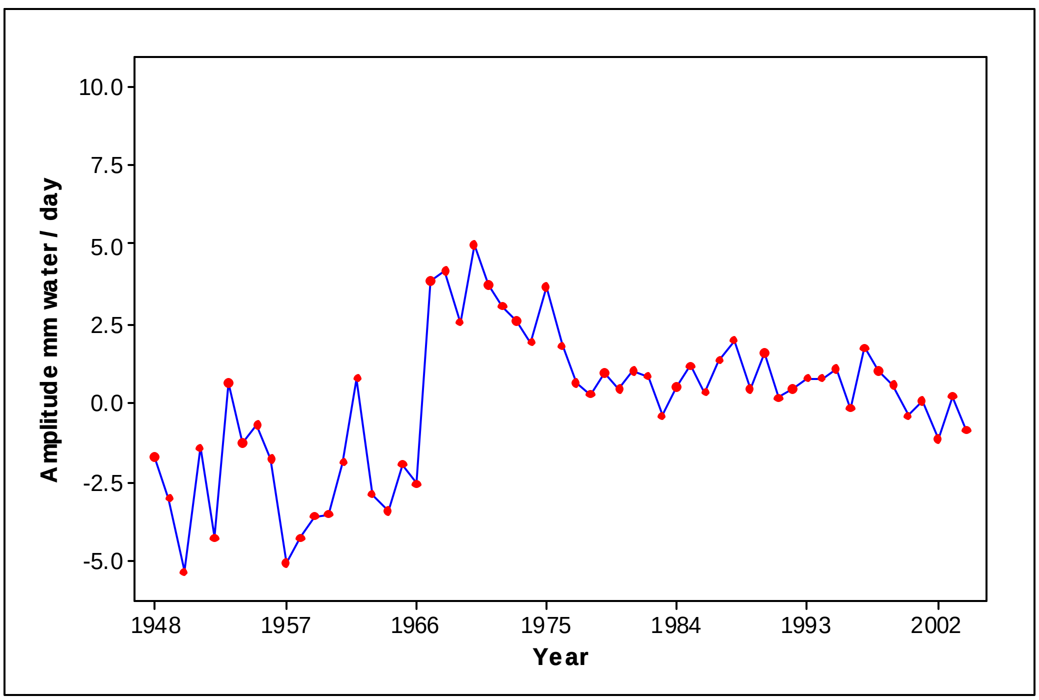
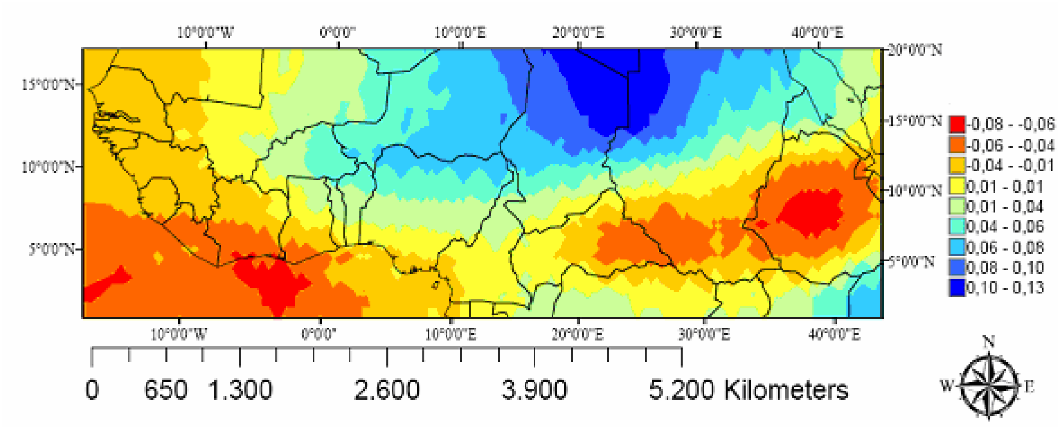


Figure 38: (a) Spatial loading of the third principal component from PCA analysis of precipitation over Central Africa (b) corresponding annual time series of factor scores of PC3

The fourth principal component represents 4.9% of the variance. Figure 38a shows its spatial eigenvector pattern. Positive loadings are observed from the mid-western through to the northern parts of the study area and the mid northern parts having positive loadings.

(a)



(b)

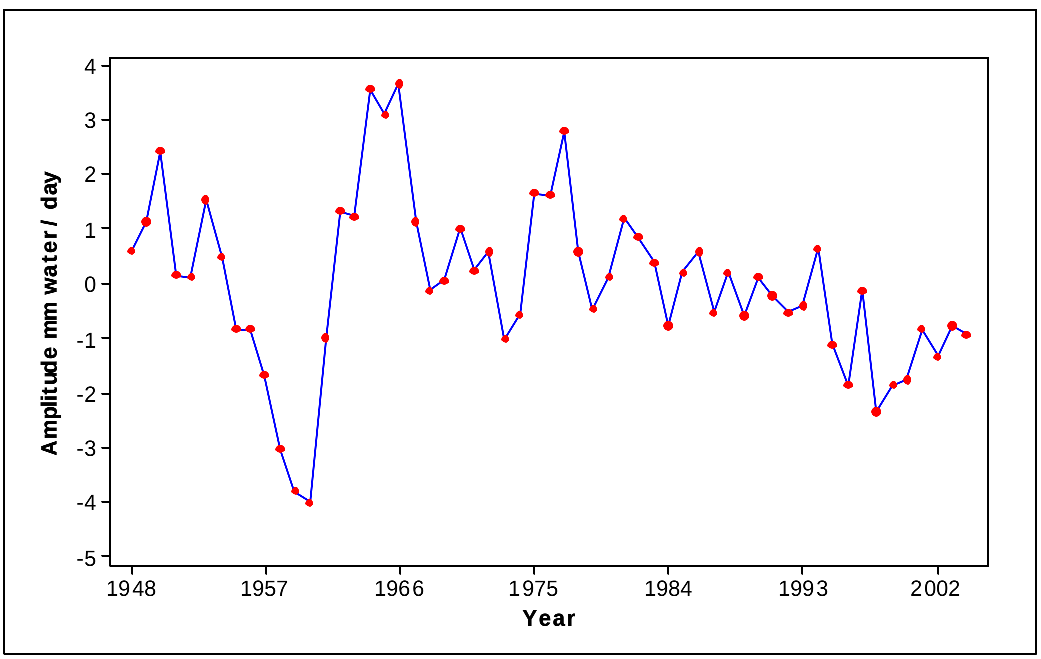


Figure 39: (a) Spatial loading of the fourth principal component from PCA analysis of precipitation over Central Africa (b) corresponding annual time series of factor scores of PC4 (c) monthly time series of the factor scores of PC4

5 Summary, Conclusion and Recommendation

This study has demonstrated that there has been a pattern continued aridity in Central Africa since the early 1970s. This pattern has been most persistent in the eastern parts of the study area particularly in countries such as Sudan and Chad. To the northern parts of the study region (the Sahel), the study reveals an increase in precipitation and decrease in temperature. What is still unclear is the cause of the increase in precipitation and temperature decrease in the Sahel, this still needs to be done. Throughout Central Africa two dry and wet periods were found in the 57 years. The wet period spans from 1949-1970 and the dry period from1971-2004. The length of the available data does not permit to do a further investigation on this matter. This still needs to be done.

Evidence for historical dry episodes of magnitude and duration comparable with the recent Central African dry phase is generally incomplete, not widespread enough in spatial extent, and of too poor in temporal resolution to be conclusive. A combination of new and/or more accurate dating methods and historical evidence, however, could help to improve our knowledge of climate variability in Central Africa.

Principal component analysis produce has been used to decomposition of the data field into spatial eigenvectors and a temporal time series. Work in this study has demonstrated that:

a) a limited number of spatial patterns are basis for African weather,

b) coordinate system for present day climate with maximum variance contribution along the first axis, maximum of the remaining variance along the second axis with subsequent axes explaining less variance can be used to explain precipitation variability within Central Africa. The same method can be used to explain temporal differences between climate change scenarios and present day climate. However this is beyond the scope of this thesis and remains to be done.

Appendix

Change in temperature and precipitation at grid point averages in Central Africa

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
-16.75	18	0.009	0.002
-14.25	18	0.004	0.004
-11.75	18	-0.006	0.005
-9.25	18	-0.013	-0.003
-6.75	18	-0.019	-0.010
-4.25	18	-0.023	-0.022
-1.75	18	-0.027	-0.033
1	18	-0.031	-0.079
3.25	18	-0.033	-0.122
5.75	18	-0.035	-0.138
8.25	18	-0.036	-0.178
10.75	18	-0.038	-0.232
13.25	18	-0.039	-0.251
15.75	18	-0.038	-0.319
18.75	18	-0.032	-0.382
20.75	18	-0.025	-0.240
23.25	18	-0.018	-0.171
25.75	18	-0.009	-0.128
28.25	18	-0.002	-0.114
30.75	18	0.005	-0.121
33.25	18	0.015	-0.176
35.75	18	0.021	-0.292
38.25	18	0.019	-0.286
40.75	18	0.018	-0.242
43.25	18	0.025	-0.216
45.75	18	0.030	-0.131

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
48.25	18	0.027	-0.022
50.75	18	0.020	-0.006
-16.75	15.5	0.004	-0.203
-14.25	15.5	-0.001	-0.186
-11.75	15.5	-0.009	-0.092
-9.25	15.5	-0.018	-0.116
-6.75	15.5	-0.021	-0.174
-4.25	15.5	-0.020	-0.274
-1.75	15.5	-0.018	-0.303
1	15.5	-0.015	-0.399
3.25	15.5	-0.015	-0.535
5.75	15.5	-0.014	-0.691
8.25	15.5	-0.013	-0.761
10.75	15.5	-0.013	-0.721
13.25	15.5	-0.014	-0.654
15.75	15.5	-0.013	-0.654
18.75	15.5	-0.008	-0.912
20.75	15.5	0.001	-0.428
23.25	15.5	0.008	-0.405
25.75	15.5	0.013	-0.326
28.25	15.5	0.019	-0.298
30.75	15.5	0.029	-0.316
33.25	15.5	0.042	-0.424
35.75	15.5	0.046	-0.644
38.25	15.5	0.037	-0.579
40.75	15.5	0.027	-0.367
43.25	15.5	0.031	-0.358
45.75	15.5	0.031	-0.225
48.25	15.5	0.029	-0.022

Longitude	**Latitude**	**Temperature change K/year**	**Precipitation change mm/year**
50.75	15.5	0.022	-0.020
-16.75	13	0.007	-0.156
-14.25	13	0.002	-0.144
-11.75	13	-0.008	-0.081
-9.25	13	-0.018	-0.122
-6.75	13	-0.018	-0.238
-4.25	13	-0.015	-0.457
-1.75	13	-0.011	-0.493
1	13	-0.003	-0.522
3.25	13	0.001	-0.641
5.75	13	0.001	-0.893
8.25	13	0.008	-1.035
10.75	13	0.010	-1.018
13.25	13	0.012	-0.846
15.75	13	0.017	-0.843
18.75	13	0.025	-0.918
20.75	13	0.030	-1.240
23.25	13	0.030	-1.325
25.75	13	0.027	-1.050
28.25	13	0.030	-0.989
30.75	13	0.040	-0.987
33.25	13	0.050	-0.922
35.75	13	0.051	-0.944
38.25	13	0.042	-0.827
40.75	13	0.035	-0.491
43.25	13	0.031	-0.296
45.75	13	0.028	-0.185
48.25	13	0.024	-0.048
50.75	13	0.020	-0.034

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
-16.75	10.5	0.012	-0.342
-14.25	10.5	0.012	-0.303
-11.75	10.5	0.007	-0.159
-9.25	10.5	0.000	-0.230
-6.75	10.5	0.002	-0.343
-4.25	10.5	0.001	-0.533
-1.75	10.5	0.004	-0.557
1	10.5	0.016	-0.590
3.25	10.5	0.021	-0.735
5.75	10.5	0.022	-0.952
8.25	10.5	0.021	-0.993
10.75	10.5	0.020	-0.881
13.25	10.5	0.021	-0.899
15.75	10.5	0.026	-1.068
18.75	10.5	0.032	-1.439
20.75	10.5	0.034	-1.711
23.25	10.5	0.029	-1.666
25.75	10.5	0.023	-1.279
28.25	10.5	0.026	-1.415
30.75	10.5	0.035	-1.455
33.25	10.5	0.041	-1.167
35.75	10.5	0.037	-1.007
38.25	10.5	0.032	-0.927
40.75	10.5	0.031	-0.674
43.25	10.5	0.030	-0.248
45.75	10.5	0.026	-0.095
48.25	10.5	0.022	-0.049
50.75	10.5	0.019	-0.031
-16.75	8	0.013	-0.376

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
-14.25	8	0.016	-0.343
-11.75	8	0.019	-0.158
-9.25	8	0.020	-0.164
-6.75	8	0.018	-0.157
-4.25	8	0.017	-0.120
-1.75	8	0.016	-0.145
1	8	0.028	-0.355
3.25	8	0.033	-0.555
5.75	8	0.032	-0.635
8.25	8	0.027	-0.598
10.75	8	0.023	-0.460
13.25	8	0.021	-0.518
15.75	8	0.022	-0.693
18.75	8	0.025	-1.133
20.75	8	0.026	-1.263
23.25	8	0.026	-1.188
25.75	8	0.024	-1.092
28.25	8	0.027	-1.438
30.75	8	0.032	-1.535
33.25	8	0.032	-1.216
35.75	8	0.026	-0.984
38.25	8	0.020	-0.888
40.75	8	0.021	-0.653
43.25	8	0.024	-0.274
45.75	8	0.024	-0.125
48.25	8	0.022	-0.094
50.75	8	0.020	-0.069
-16.75	5.5	0.014	-0.133
-14.25	5.5	0.015	-0.164

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
-11.75	5.5	0.017	-0.150
-9.25	5.5	0.020	0.108
-6.75	5.5	0.019	0.270
-4.25	5.5	0.016	0.331
-1.75	5.5	0.013	0.195
1	5.5	0.018	-0.326
3.25	5.5	0.021	-0.690
5.75	5.5	0.021	-0.537
8.25	5.5	0.019	-0.300
10.75	5.5	0.020	-0.004
13.25	5.5	0.020	-0.115
15.75	5.5	0.021	-0.272
18.75	5.5	0.021	-0.601
20.75	5.5	0.021	-0.733
23.25	5.5	0.022	-0.730
25.75	5.5	0.022	-0.787
28.25	5.5	0.025	-1.095
30.75	5.5	0.029	-1.200
33.25	5.5	0.026	-0.966
35.75	5.5	0.020	-0.608
38.25	5.5	0.015	-0.480
40.75	5.5	0.015	-0.285
43.25	5.5	0.017	-0.096
45.75	5.5	0.018	-0.075
48.25	5.5	0.018	-0.135
50.75	5.5	0.017	-0.151
-16.75	3	0.016	-0.266
-14.25	3	0.015	0.009
-11.75	3	0.015	0.169

Longitude	**Latitude**	**Temperature change K/year**	**Precipitation change mm/year**
-9.25	3	0.015	-0.020
-6.75	3	0.015	-0.026
-4.25	3	0.013	0.049
-1.75	3	0.011	0.094
1	3	0.009	-0.691
3.25	3	0.008	-0.020
5.75	3	0.009	-0.941
8.25	3	0.010	-0.600
10.75	3	0.013	0.070
13.25	3	0.015	0.122
15.75	3	0.017	-0.027
18.75	3	0.016	-0.207
20.75	3	0.015	-0.226
23.25	3	0.014	-0.277
25.75	3	0.015	-0.322
28.25	3	0.017	-0.486
30.75	3	0.021	-0.631
33.25	3	0.021	-0.651
35.75	3	0.018	-0.350
38.25	3	0.013	-0.169
40.75	3	0.011	0.017
43.25	3	0.011	0.035
45.75	3	0.012	-0.020
48.25	3	0.013	-0.152
50.75	3	0.014	-0.135
-16.75	0.5	0.013	0.120
-14.25	0.5	0.013	0.098
-11.75	0.5	0.014	0.070
-9.25	0.5	0.014	0.093

Longitude	Latitude	Temperature change K/year	Precipitation change mm/year
-6.75	0.5	0.015	0.090
-4.25	0.5	0.015	-0.022
-1.75	0.5	0.014	-0.217
1	0.5	0.012	-0.413
3.25	0.5	0.010	-0.569
5.75	0.5	0.009	-0.689
8.25	0.5	0.012	-0.548
10.75	0.5	0.014	-0.050
13.25	0.5	0.014	0.180
15.75	0.5	0.013	0.107
18.75	0.5	0.012	-0.051
20.75	0.5	0.012	0.005
23.25	0.5	0.012	-0.042
25.75	0.5	0.012	-0.103
28.25	0.5	0.012	-0.164
30.75	0.5	0.013	-0.240
33.25	0.5	0.015	-0.247
35.75	0.5	0.015	-0.126
38.25	0.5	0.012	-0.020
40.75	0.5	0.010	0.058
43.25	0.5	0.009	0.036
45.75	0.5	0.011	-0.015
48.25	0.5	0.011	-0.089
50.75	0.5	0.012	-0.038

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Glossary

Acronyms

AEJ	African Easterly Jetstream
AEW	African Easterly Waves
BP	Before Present
CRU	Climate Research Unit
EOF	Empirical orthogonal functions
GHG	Greenhouse Gas
GCOS	Global Climate Observing system
ITCZ	Inter Tropical Convergence Zone
LGM	Last Glacial Maximum
MAPE	Mean Absolute Percentage Error
MAD	Mean Absolute Deviation
MSLP	Mean Sea Level Pressure
NADW	North Atlantic Deep Water
NCAR	National Centre for Atmospheric Research
NCEP	National Centres for Environmental Prediction
NDVI	Normalised Difference Vegetation Index
NOAA	National Oceanic and Atmospheric Administration
NPP	Net Primary Production
OGCM	Ocean General Circulation Model
PCA	Principal Component Analysis
PC	Principal Components

RUE	Rainfall use efficency
SAT	Surface Air Temperature
SN	Signal to Noise Ratio
SST	Sea Surface Temperature
SSTA	Sea Surface Temperature Anomalies
ST	J Subtropical Jetstream
TEJ	Tropical Easterly Jet (TEJ)
WAM	West African Monsoon

Mathematical symbols

b	Regression amplitude estimator
a	Constant factor estimator
∂	Partial differential
Δ	A small change
k	Truncation at which the regression is performed
f	Scalar linear diagnostic of the climatic system
s	Standard deviation
x^2	Chi-square distribution
$\wedge$	Estimate of the variable
$\sum$	Summation operator
I	Identity matrix
r^2	Sum of residual squares
X^T	Transpose of X

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